THE UNIVERSITY OF ALBERTA

RELEASE FORM

NAME OF AUTHOR Richard D. Lamoreaux

TITLE OF THESIS Cluster Patterns in Seismicity

DEGREE FOR WHICH THESIS WAS PRESENTED Doctor of Philosophy

YEAR THIS DEGREE GRANTED Spring 1982

THE UNIVERSITY OF ALBERTA

Cluster Patterns in Seismicity

by

© Richard D. Lamoreaux

A THESIS

SUBMITTED TO THE FACULTY OF GRADUATE STUDIES AND RESEARCH

IN PARTIAL FULFILMENT OF THE REQUIREMENTS FOR THE DEGREE

OF Doctor of Philosophy

IN

Geophysics

Department of Physics

EDMONTON, ALBERTA

Spring 1982

THE UNIVERSITY OF ALBERTA

FACULTY OF GRADUATE STUDIES AND RESEARCH

The undersigned certify that they have read, and recommend to the Faculty of Graduate Studies and Research, for acceptance, a thesis entitled Cluster Patterns in Seismicity, submitted by Richard D. Lamoreaux in partial fulfilment of the requirements for the degree of Doctor of Philosophy in Geophysics.

Dedication

To my parents who have wondered what I have been doing for all these long years, but have never failed to give me less than their complete support.

Abstract

Several aspects of the clustering of earthquakes related to precursory seismicity patterns are examined. Using a simple definition for an aftershock, a self-similarity or scaling law in mainshock - aftershock clustering is proposed. The scaling law suggests that number of aftershocks does not depend on the magnitude of the mainshock if the aftershocks are counted in a fixed magnitude range below that of the mainshock. This scaling law reduces the number of observations necessary for statistical studies of aftershock sequences and imposes a constraint on acceptable physical models of aftershocks. On the average, foreshocks seem to generate larger numbers of aftershocks than do mainshocks. The scaling law supports the hypothesis that the distribution of irregularities in material strengths on a fault zone shows no characteristic scale.

The possibility of local clustering in earthquake sequences in Southern California with longer time and distance scales than the obvious "mainshock-aftershock" clustering is explored. To study the possibility of long-term clustering "foreshocks" and "aftershocks" are removed from the catalogue using a simple box car filter definition for the foreshocks and aftershocks. The lengths of the box car filters were dependent on the magnitude of the foreshock and mainshock. It is demonstrated that even if foreshocks and aftershocks are elimiminated through such a filtering process, the residual catalogue still shows weak

but significant clustering characteristics. This clustering is evident on time scales of the order of 3 to 6 years and distance scales of the order of 100 kilometers for residual events ("mainshocks") with magnitudes greater than or equal to 4. This observation supports suggestions that sequences of earthquakes are interdependent over longer times and distances than is widely accepted. The possibility of longer period clustering with periods greater than 6 years could not be explored due to the short length of the catalogue.

Anomalous periods of clustering and anti-clustering in the sequence of residual events were juxtaposed with the occurrences of large events in the same region. Anomalous groups of residual events, swarms or concentrated clusters of "mainshocks", preceded by quiescence were found to precede, most but not all, large earthquakes in the Southern California region. The quiescence → activation pattern observed here for mainshocks is similar to precursory patterns observed by other investigators, although the time scale is in general longer. The mathematical technique of cluster analysis is shown to be useful in the study of anomalous seismicity patterns.

Three proposed precursory seismic patterns termed "bursts of seismicity" are applied to retrospective prediction of large earthquakes which have occurred in the vicinity of the Cocos - North American - Caribbean triple junction. The three patterns, "bursts of aftershocks", swarms, and sigma, are based on abnormal clustering of

earthquakes in time, space and energy. The patterns give respectable results, although the available data set is not of the highest quality. All occurrences of the patterns are close to the proposed location of the unstable triple junction. Success of the patterns may be due to the complex nature of the tectonics in the triple junction area.

Acknowledgements

I would like to thank my supervisor Dr. Edo Nyland for his support, suggestions, and gentle prods throughout the course of this study. I would also like to thank Dr. Nyland for his part in the arrangement of my period of study with Professor V. I. Keilis-Borok.

To Prof. Keilis-Borok I owe many thanks. It was my extreme good fortune to be able to work under the direction of Prof. Keilis-Borok for a short period of time. Most of the work in this thesis was a direct result of my studies with Prof. Keilis-Borok. Thank you Volodja for your patience and understanding as a teacher. You have had a great influence on my outlook as a scientist. I greatly value the time I spent working with you.

I am grateful to the Seismological Labotatory of the California Institute of Technology for the support I recieved during my studies there. L. Lenches drafted many of the figures given in Chapter 3.

Without the generous support of the Dept. of Physics at the University of Alberta in the form of teaching assistantships, travel advances, and a dissertation fellowship this work would not have been possible.

The excellent computing facilities of the University of Alberta were invaluable in this study.

I would like to thank my good friends Dean Mackay, Gary Margrave, William Tighe, and Delia Todd for many good times and useful discussions. The frequent trips we made to the

mountains preserved my sanity and made this one of the most enjoyable periods of my life.

I would like to thank Linda Hall for the donating a considerable amount of her time in the editing of this work. Thank you Linda, you made a difficult task bearable.

Last but not least, I would like to thank my parents. Without your unselfish financial and emotional support thoughout my studies none of this would have been possible.

Table of Contents

List of tables

anomalous for $\Delta = 2.0$ and $\bar{B}(q) = 90\%$.

Table 3.1 (page 78) gives a simplified version of R(M) and T(M) used in the identification of aftershocks (after Keilis-Borok et al., 1980b).

Table 3.2 (page 79) is a tabulation of the number of mainshocks in different time and magnitude intervals. Mainshocks are defined by table 2.1. Each column indicates the number of events from the magnitude interval heading the column to but not including the magnitude heading the next column. The time interval is indicated with the last year not inclusive. The average magnitude is calculated for magnitudes greater than 3.5. The stability of the average magnitude is an indication of catalogue completeness for magnitudes above 3.5.

Table 3.3 (page 80) gives the same tabulation as table 3.2 but uses the mainshock definition of table 3.1

Table 3.4 (page 85) gives an example of P(n,N) using the mainshock definition of table 2.1 and mainshocks with $M \geq 3.5$.

Table 3.5 (page 86) gives an example of P(n,N) using the mainshock definition of table 3.1 and mainshocks with $M \geq 3.5$.

Table 3.6 (page 87) gives a Chi-squared comparison of the two dimensional histograms to the corresponding distributions predicted by a Poisson process. Foreshocks are included.

Table 3.7 (page 89) gives a Chi-squared comparison of the two dimensional histograms P(n,N), excluding foreshocks, to Poisson distributions.

Table 3.8 (page 98) gives a listing of large events, $M \geq 6.4$, in Southern California.

Table 3.9 (page 126) gives a tabulation of the centroids for clusters found in the northern region of Southern California for the case $M_0=4.0$ and 50 km. equal to 3 years (i.e. $t_0=3$ years, $x_0=.548393°$ and $y_0=.449507°$).

Table 3.10 (page 127) gives a tabulation of the centroids for clusters found in the southern region of Southern California for the case $M \geq 4.0$ and 30 km. equal to 2 years (i.e. $t_0=2$ years, $x_0=.329036°$ and $y_0=.269704°$).

Table 4.1 (page 158) gives the numbers of events listed in the NOAA catalogue in various magnitude ranges of $\bar{M}$ for five year intervals starting in 1900. The table includes events in the entire Middle America region.

Table 4.2 (page 161) gives a list of large ($M \geq 7.1$) events occurring in the study region since 1920.

Table 4.3 (page 167) gives a listing of all mainshocks ($M \geq 6.0$) with 4 or more aftershocks in the 2 day period following the mainshock. Also listed are the large ($M \geq 7.2$) earthquakes which have occurred in the region since 1950.

List of figures

values of Δ. The distributions are based on all mainshocks for which the catalogue was considered sufficiently complete (ie $Mm \geq m_0(t) + \Delta$, where $m_0(t)=3.5$ until 1970, 3.0 from 1970 to 1977 and 2.5 after 1977). The distributions for $\Delta=2.0$ and $\Delta=2.5$ are based on small numbers of points (64 and 31 respectively).

Figure 2.6 (page 58) give a comparison of the normalized cumulative frequency distributions b(q) for the volcanic and the non-volcanic regions ($\Delta=1.5$).

Figure 2.7 (page 64) shows the normalized cumulative frequency distributions b(q) for foreshocks and mainshocks with $\Delta=1.0$. Aftershocks were counted for 2 days.

Figure 3.1 (page 81) shows a map of mainshocks for Southern California with magnitudes greater than 5.0. The aftershock thresholds of table 2.1 were used. The boundaries of the 14 subregions are also given.

Figure 3.2 (page 82) gives a plot of the cumulative number of mainshocks Ns, $M\geq4.0$, from the beginning of the catalogue.

Figure 3.3 (page 100) gives the test of the quiescence $\rightarrow$ activation hypothesis for the Owens Valley and Imperial Valley regions.

Figure 3.4 (page 101) gives the test of the quiescence $\rightarrow$ activation hypothesis for the Santa Barbara and Ensenada regions.

Figure 3.5 (page 102) gives the test of the quiescence $\rightarrow$ activation hypothesis for the Southern Mojave and Laguna

Salada regions.

Figure 3.6 (page 103) gives the test of the quiescence → activation hypothesis for the Parkfield region.

Figure 3.7 (page 105) gives the test of the quiescence → activation hypothesis for the San Jacinto and Kern County regions.

Figure 3.8 (page 106) gives the test of the quiescence → activation hypothesis for the Southern Sierra and Los Angeles Basin regions.

Figure 3.9 (page 107) gives the test of the quiescence → activation hypothesis for the Riverside and Elsinore regions.

Figure 3.10 (page 108) gives the test of the quiescence → activation hypothesis for the Northern Mojave region.

Figure 3.11 (page 110) gives a rough summary of the results of of part 1. Periods of quiescence are indicated by dark circles and swarms are indicated by open circles. The minimal magnitude for the region and large earthquakes occurring in the region are also indicated.

Figure 3.12 (page 128) gives the locations of the cluster centroids listed in tables 3.9 and 3.10. The boundary between the northern and southern regions is indicated by the dashed line.

Figure 3.13 (page 131) gives a time magnitude plot of earthquakes in the Owens Valley region (figure 3.1). Earthquakes not found to be cluster members are indicated by a square symbol. The times of the clusters are indicated by

horizontal bars.

Figure 3.14a (page 132) gives latitude longitude plot of the clusters found in the Owens Valley region.

Figure 3.14b (page 133) gives a continuation of Figure 3.14a.

Figure 3.15 (page 135) gives a latitude longitude plot of the clusters found in the the Southern Sierra region.

Figure 3.16a (page 137) gives a latitude longitude plot of clusters found in the San Jacinto region.

Figure 3.16b (page 138) gives a continuation of 3.16a

Figure 3.16c (page 139) gives a continuation of 3.16a and 3.16b.

Figure 3.17 (page 141) gives a time magnitude plot for the members of the clusters in the San Jacinto region. The times of the clusters are indicated by horezontal bars.

Figure 3.18 (page 144) gives a contour plot of the average $\hat{\delta}$ for the Southern California region using sliding 6 year intervals stepped one year. Anomalous values of $\hat{\delta}$ have been eliminated from the average.

Figure 3.19 (page 145) gives a map of $\hat{\delta}$ for the years 1945 to 1951.

Figure 3.20 (page 146) gives a map of $\hat{\delta}$ for the years 1950 to 1956.

Figure 3.21 (page 147) gives a map of $\hat{\delta}$ for the years 1975 to 1981.

Figure 4.1 (page 159) gives an outline of the area used in the bursts of seismicity study along with large

earthquakes ($M \geq 6.0$) listed in the NOAA catalogue for the area.

Figure 4.2 (page 166) shows plots of $n(t)/\tilde{N}(t)$ and $r(t)/n(t)$ for the periods 1932 to 1959 (a) and 1952 to 1980 (b). Large earthquakes ($M \geq 7.2$) are indicated by vertical lines. Dashed lines indicate slightly smaller events of events occurring just outside the study region.

Figure 4.3 (page 172) shows plots of $\mathbb{S}$ for the periods 1932 to 1960 (a) and 1957 to 1980 (b). Large events are indicated as in Figure 4.2.

1. Introduction

1.1 Preliminary remarks

Variations in seismicity patterns before large earthquakes have been studied by many investigators. These studies serve dual purposes. Very little is known about the physical mechanism of earthquake failure and in particular about the nature of any inter-relationships between two or more earthquakes in a given region. Premonitory seismicity patterns place constraints on acceptable physical models which attempt to explain these phenomena. Also, seismicity patterns may be useful for earthquake prediction.

In general the study of seismicity patterns can be approached differently depending on the purpose of the study. If the purpose is to develop an algorithm which is directly useful for earthquake prediction it is necessary to treat the seismicity patterns in as rigorous a manner as is statistically possible. Completeness and homogeneity of earthquake catalogues, and uniformity of magnitude estimates are critical. Most studies have searched for anomalous seismicity before individual events without attempting to generalize the observed pattern. In this approach rigorous definitions of seismicity patterns and uniformity of the catalogue are less important. As a result most observed precursory seismicity patterns are only loosely defined.

Precursory seismicity patterns can be grouped into three basic categories, quiescence (including seismic gaps), activation (including swarms and foreshocks), and migration. Large earthquakes, at least those along simple plate boundaries, apparently repeat with intervals of a few tens to hundreds of years (for example Rikitake, 1976a). A number of investigators have tried to find periodities (recurrence intervals) and regularities in their occurrence. The main content of this chapter will be a review of observed precursory patterns and a review of studies of recurrence intervals for large earthquakes.

An important feature of earthquake occurrence with great relevance to the failure process is the apparent self similarity or scale invariance of earthquake catalogues (Kagan and Knopoff, 1980a). An aspect of this scale invariance is studied in chapter 2, a self similarity in the occurrence of aftershocks. Chapter 3 studies the clustering of mainshocks in Southern California. A precursory pattern, based on the quiescence and clustering of mainshocks, premonitory to large ($M \geq 6.4$) earthquakes in the region is proposed. Cluster analysis is applied to the study of the mainshock clustering. Chapter 4 applies the "bursts of seismicity" patterns proposed by Keilis-Borok et al. (1980b) to retrospective prediction of earthquakes in the region of the Cocos - North American - Caribbean triple junction.

1.2 Implicit assumptions

The direct use of premonitory patterns in earthquake prediction rests on several postulates. Foremost is the postulate that earthquakes in a catalogue do not follow a Poisson distribution. The occurrence of aftershocks is the most prominent deviation from Poisson behavior. With aftershocks removed there is some question as to whether the residual catalogue follows a Poisson distribution (Gardner and Knopoff, 1974). There is amibiguity in the exact definition of aftershocks. The question of longer period clustering with the obvious short term clustering removed is explored in Chapter 3. If the residual catalogue after short term clustering is removed is Poisson in character there is little hope of finding useful precursory seismicity patterns.

Other postulates concern the nature of the failure process. The site of a large earthquake will not be physically isolated from surrounding regions. Presumably the state of stress, which occurs in the epicentral area of a major earthquake prior to failure, will have a significant effect on the state of stress in surrounding regions, possibly to rather large distances. Since seismicity results from a release of accumulated strain; it seems reasonable that the nature of the seismicity in a region should depend on the amount of unreleased strain in that region. In particular, the seismicity when the average unreleased strain is high might be expected to show some non-random

differences from the seismicity when the average unreleased strain is low. Presumably, the highest average unreleased strains in a region will be achieved just prior to the occurrence of a large earthquake in that region.

In a given region large magnitude earthquakes are rare and the time spans of earthquake data sets are in general short compared to the return times for great earthquakes. This means that the statistical basis of any phenomenon associated with large magnitude earthquakes will at best be very weak for a given area. The direct use of premonitory patterns in earthquake prediction will likely fail unless there are systematic precursory patterns which do not vary substantially in form from event to event and region to region.

1.3 Some definitions

The problem of searching for premonitory patterns can be summed up as a search for patterns in the background seismicity that precede the largest earthquakes ($M \geq M_0$) in a region. The background seismicity is defined as all earthquakes with magnitudes less than M_0 and greater than some minimum threshold. The minimum threshold is usually determined by the detection capabilities of the seismic network being used. Aftershocks of the largest events are excluded from the background seismicity.

Some rules are needed to analyze the success of a given pattern. If a pattern is valid then the occurrences of the pattern should show a marked tendency to fall within some time window preceding the occurrences of the major events. The length of this time window may depend on several parameters such as the regional rate of strain accumulation, the minimum magnitude of the major events, and the type of pattern being considered. Call the length of this precursory time window t_0. A success for a pattern is then defined as the occurrence of the pattern which is followed by a major event within time t_0. A failure is defined as a major event which was not preceded by the occurrence of the pattern within time t_0. A false alarm is defined as an occurrence of a pattern which is not followed by a major event within time t_0.

1.4 Seismicity patterns and plate tectonics

One of the most prominent features of world seismicity is that the vast majority of earthquakes occur along narrow, well defined, interconnecting belts. This feature was explained by the global plate tectonic theory. The narrow seismic belts coincide with major plate boundaries. Relatively little deformation occurs in most plate interiors.

The majority of large ($M \geq 7.5$) earthquakes occur at shallow depths along either convergent boundaries where

active subduction of one lithospheric plate is occurring or along strike-slip type boundaries. Large events account for most of the total seismic slip, and seismic energy release for these types of boundaries (Brune, 1968).

1.4.1 Seismic Gaps

Presumably strain energy is gradually accumulated at the plate boundaries due to the relative motion between plates. Accordingly, the largest earthquakes in the same region will usually be separated by many years. It has been shown that rupture zones for large earthquakes do not overlap appreciably (Mogi, 1968a; Sykes, 1971). Since a large portion of the relative plate motion along convergent and strike slip boundaries is accounted for by the slip during major earthquakes, it follows that, over a suitably long period of time, rupture zones of the largest earthquakes will tend to cover a seismic zone. Gaps in the spatial distribution of these rupture zones are then likely locations for large earthquakes in the future. These gaps are termed seismic gaps or seismic gaps of the first kind (Mogi, 1979). The seismic gap hypothesis can be summed up as follows: segments of major plate boundaries that have not been the sites of large earthquakes for tens to hundreds of years are more likely to be the sites of future large earthquakes than segments that have experienced major rupture more recently.

Seismic gaps were first observed by Fedotov (1965) and successfully applied in the Kamchatka, Kurile Island and northeastern Japanese seismic regions. Mogi (1968a) independently applied the pattern to the Japanese and Alaska seismic zones, noting that gaps in seismic activity in the Japanese seismic zone have been successively filled within several tens of years by great earthquakes without significant overlap of their rupture zones. Subsequent papers (Sykes, 1971; Kelleher, 1970, 1972; Kelleher et al., 1973; Ohtake et al., 1977; McCann et al., 1979) have applied the technique with great success to other seismic regions. McCann et al. (1979) gives a comprehensive review.

The seismic gap technique provides estimates of the location and maximum likely size for some future large earthquakes, however, it does not constrain the time of occurrence to closer than decades (McCann et al., 1979). Another complication is the apparent existence of permanent seismic gaps (Kelleher and McCann, 1976,1977). Kelleher and McCann (1976,1977) found, contrary to what would be expected from plate tectonic considerations, sections of some island arcs have experienced very infrequent or no large earthquakes in the time spanned by present earthquake data. Most of these zones are at or near regions where groups of seamounts, aseismic ridges or other topographic features on the subducting plate intersect the subduction zone (Kelleher and McCann, 1976). They propose that in these regions the subduction process is modified due to the relative buoyant

material associated with the topographic features. The seismic gap method is not applicable in these regions (McCann et al., 1979).

1.4.2 Frequency laws

Most studies of seismic risk begin with an estimate of the frequency of occurrence of earthquakes with different magnitudes in the region of interest. These estimates are usually based on the historical records of seismicity. Unfortunately, historical records are too short to evaluate the frequency of occurrence with any confidence. Molnar (1979) suggests that intervals of the order of 10^3 years in most places and 10^4 years in some places are required.

Gutenberg and Richter (1954) and others have shown that the number of earthquakes, N(M), with magnitudes greater than or equal to M is approximated well by

$$\log_{10} N(M) = a - bM \qquad 1$$

where a and b are constants for a given region. Due to the short lengths of time spanned by earthquake data sets, the constants a and b probably can not be accurately determined unless large regions are considered. If large regions are taken the analysis loses much of its usefulness.

A useful parameter to characterize the size of an earthquake is the seismic moment:

$$M_0 = \gamma u_0 A \qquad 2$$

where γ is the shear modulus, A is the fault area and u_o is the average slip on the fault during the earthquake. The summation of the seismic moments for individual earthquakes occurring on a fault can be used to estimate the average slip on a fault (Brune, 1968). Let L be the length of the fault, and W the width. Brune (1968) showed that the average displacement on the fault $\bar{U}_o$ after a sufficiently long period of time is given by

$$\bar{U}_0 = \Sigma (M_0 / \gamma LW) \qquad 3$$

where the sum is over all earthquakes on the fault. The above assumes that no motion occurs by fault creep. Average rates of slip estimated tectonically or otherwise can be used to regionally constrain the frequency of occurrence of earthquakes with different magnitudes. Using this constraint, equation 3, and the Gutenberg Richter frequency magnitude relation Molnar (1979), derives a frequency of occurrence law with more local applicability.

$$N(M_0) = \frac{(1-\beta)\bar{\gamma} A \bar{v} M_0^{-\beta}}{M_0(max)^{(1-\beta)}} \qquad 4$$

$N(M_o)$ is the number of events with seismic moments greater than or equal to M_o, $\bar{\gamma}$ is the average shear modulus on the fault zone, A is the area of the fault zone, $\bar{v}$ is the average slip rate excluding creep, β is a constant which can be determined independently and $M_o(max)$ is the maximum possible seismic moment for an event in the region. Anderson

(1979) uses a similar method with geological estimates of regional strain rates as the constraint. With only presently available data there are large uncertainies involved in estimating the values of the above parameters (β, $\bar{\gamma}$, $\bar{v}$, etc.). Equation 4 probably results in only a modest improvement over estimates which use historical data alone (Molnar, 1979). The Gutenberg Richter relationship may not be valid for largest magnitude events in some regions (Singh et al., 1981). If this is the case the methods of Molnar (1979) and Anderson (1979) are not valid in these zones (Singh et al., 1981).

1.4.3 Recurrence intervals

Equations 1 and 4 are, at present, probably only useful in estimating mean seismicity patterns for large regions. Estimates of recurrence intervals (repeat times) for individual fault segments would be more useful. The recurrence of large earthquakes which rupture nearly the same portion of a plate boundary have long been documented even before the introduction of plate tectonics. Early investigation of historical Japanese data (Imamura, 1928) indicated that large earthquakes in southwest Japan occurred repeatedly, at approximately the same location, with repeat times of 100 to 150 years. Similar regularities have been report in other regions by many authors (for example, Mogi, 1968a; Fedotov et al., 1970; Kelleher, 1970, 1972; Sykes, 1971; Utsu, 1974; Rikitake, 1976a; Kelleher et al., 1973,

1974; Ando, 1975; Sieh, 1978; McNally and Minster, 1981; Singh et al., 1981; Sykes and Quittmeyer, 1981).

The short time span of historical records for most regions make the statistical determination of repeat times at a location unreliable. For simple plate boundaries average repeat times ($\bar{T}$) can be estimated by

$$\bar{T} = \bar{u}/\alpha v \qquad 5$$

where $\bar{u}$ is the average displacement which occurs seismically at the location, v is the relative plate velocity, and α is the ratio of seismic slip to total slip (Sykes and Quittmeyer, 1981). The above assumes that relative plate velocities are uniform over long periods of time (Sykes and Quittmeyer, 1981), and that the majority of seismic slip in a region occurs during large earthquakes (Singh et al., 1981). Both assumptions seem to be reasonable (Sykes and Quittmeyer, 1981; Minster and Jordan, 1978; Brune, 1968).

Equation 5 can be used to estimate return times from the history of previous events if α is constant locally over many earthquake cyles. This requires that some focal parameters of previous events are known from which $\bar{u}$ can be estimated. For example, $\bar{u}$ can be derived from the seismic moment, M_o, using equation 2. Equation 5 then becomes $\bar{T} = M_o/\gamma A v \alpha$. Forms of equation 5 relating $\bar{T}$ to other focal parameters are given by Sykes and Quittmeyer (1981), Acharya (1979), and Singh et al. (1981).

Time intervals between successive earthquakes at a given location are irregular enough that for most regions a knowledge of average recurrence interval does not greatly constrain future times of occurrence (Rikitake, 1976a; Ito, 1980; Ellsworth et al., 1981). Ito (1980) suggests that much of the irregularity in recurrence intervals could be due to interaction between seismic events.

The stress drop in an earthquake is proportional to the coseismic slip (for example, Chinnery, 1969). Using this Shimazaki and Nakata (1980) suggest a simple model for estimating recurrence intervals of individual events. Let τ_1 be the shear stress at which the earthquake initiates, and let τ_2 be the final value after the earthquake. If the rate of accumulation of stress is constant in time, the cyclic nature of stress relief associated with a sequence of earthquakes will be related to the manner in which τ_1 and τ_2 change from event to event (Shimazaki and Nakata, 1980). A strictly periodic sequence of events would result if τ_1 and τ_2 are both time independent. There will be no regularity if both τ_1 and τ_2 vary in time. If τ_2 varies with time while τ_1 remains constant a sequence of events results in which the time of occurrence of an event would depend on the amount of coseismic displacement of the preceding event. Shimazaki and Nakata (1980) call this the "time-predictable" model. If τ_2 remains constant and τ_1 varies "slip-predictable" recurrence occurs. The amount of slip of subsequent events will be proportional to the total time elapsed from the preceding

earthquake, the time of occurrence will be uncertain. The time-predictable and slip-predictable models are illustrated in figure 1.1.

Several studies find support for the time-predictable model. Bufe et al. (1977) found that the time interval between small shocks along a segment of the Calaveras fault in California depended on the displacement of previous shocks. Similar findings are reported for three sequences of earthquakes and earthquake related phenomena in Japan (Shimazaki and Nakata, 1980). Sykes and Quittmeyer (1981) found good agreement between calculated and actual recurrence times for sequences of events along many simple boundaries. They also use the time-predictable and slip-predictable models to calculate α of equation 5. The time-predictable model gave realistic values whereas the slip-predictable model did not.

If the slip-predictable model is valid the occurrence times of future large events along simple plate boundaries can be greatly constrained provided the coseismic slips of several previous events at the location are known (Sykes and Quittmeyer, 1981).

1.5 Quiescence

Quiescence refers to a spatial-temporal reduction (gap) of seismic activity in comparison to the normal activity. In the literature, quiescence is often confused with seismic

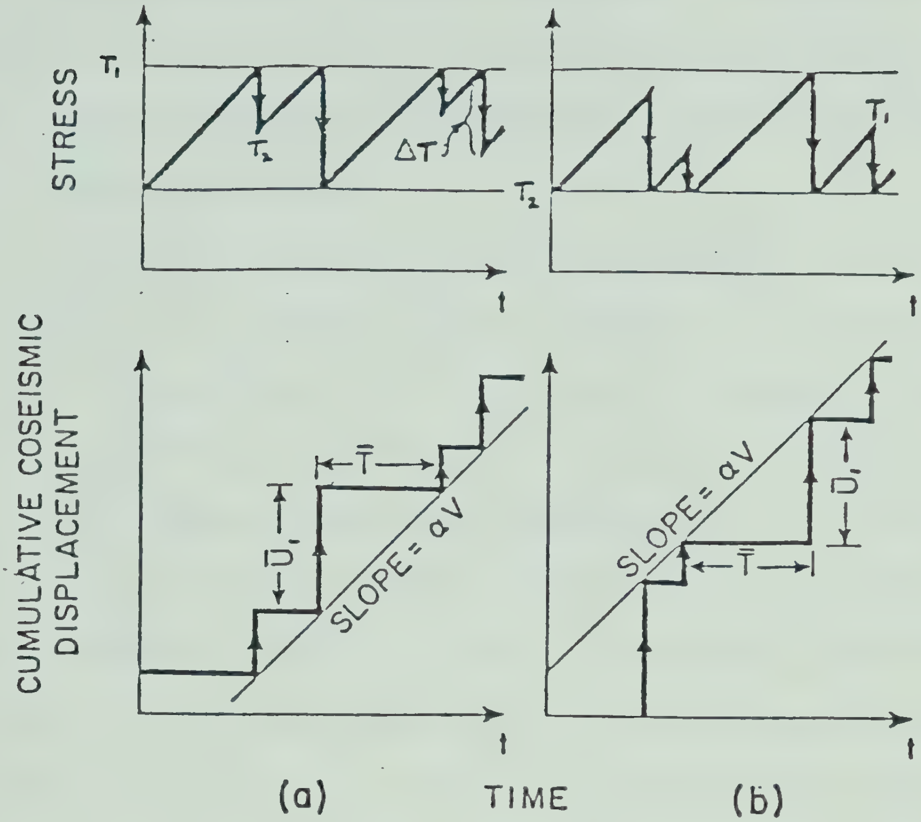


Figure 1.1 is an illustration of the "time-predictable" (a) and the "slip-predictable" (b) models of recurrence (after Shimazaki and Nakata, 1980). In model (a) the stress at which the earthquake initiates τ_1 is constant for all shocks but the final stress τ_2 varies. This results in a pattern in which the time to the next event (T)is determined by the slip ($\bar{U}_0$) of the last event. In model (b) the stress at which the earthquake initiates varies while the final stress remains constant. This results in a sequence in which the slip of the next event in proportional to the time from the last event. V is the relative plate velocity, and α is the ratio of seismic slip to total slip.

gaps. The term seismic gap will be reserved for gaps in the spatial distribution of rupture zones of the largest earthquakes in a seismic belt. Mogi (1979) refers to spatial-temporal gaps in the seismicity pattern as seismic gaps of the second kind.

A reduction of smaller magnitude seismic activity in a focal region of a future large earthquake is commonly observed. This pattern was first noted by Inouye (1965) for several large shallow earthquakes in Japan. Kelleher and Savino (1975) demonstrated that seismic gaps are also often gaps for smaller magnitude activity and that such gaps commonly persist until the time of the future large event. Other studies have reported the pattern for earthquake with magnitudes as low as 5 in virtually all tectonic settings including intraplate seismic zones (for example, Mogi, 1968a, 1969a; Borovik et al., 1971; McEvilly et al., 1967; Evison, 1977a, 1977b, 1977c; Sekiya, 1977; Engdahl and Kisslinger, 1977; Mizone et al., 1978; Kelleher and Savino, 1975; Ohtake et al., 1977; Ishida and Kanamori, 1977; Wei et al., 1978; Kristy and Simpson, 1980; Yamashina and Inoue, 1979; Kanamori, 1981; Wyss and Habermann, 1979; and Khattri and Wyss, 1978). Kanamori (1981) gives a tabulation of earthquakes for which the pattern has been observed, noting that preseismic quiescence appears to to be the most commonly observed precursory pattern.

The above studies suggest that seismic activity in the region of the eventual rupture zone of a large earthquake

decreases more or less abruptly before the major event. The nature of this decrease is poorly understood and seems to vary significantly from event to event. Mogi (1979) notes that the reduced levels of seismicity may extend only several magnitude ranges below the future large event; smaller earthquakes may or may not show normal activity. Also, for an individual event the spatial extent and time interval of the quiescence may vary significantly depending on the magnitude interval used to define the seismic activity. Ishida and Kanamori (1978) and Kanamori (1981) note that there are many quiet zones and quiescent periods which are not followed by a large earthquake. Quiescence alone is not a reliable indicator of a future large earthquake.

1.6 Activation

Activation refers to a spatial-temporal increase in the seismic activity of a region in comparison to normal activity. Activation can be manifest through increases in seismicity rates, increases in the total seismic energy release, or both. Reported increases are regional or local in nature. Swarms and foreshocks are examples of local activation.

1.6.1 Foreshocks

Foreshocks are the most obvious premonitory phenomenon preceding major earthquakes. Foreshocks are events or swarms which are close in space and time to the future mainshock. However, there is no unambiguous, widely accepted definition of what is meant by "close in space and time". Some earthquakes are preceded by very clear increases in the number of smaller earthquakes in their epicentral area a few days or weeks before they occur. Such events can unambiguously be called foreshocks. Such a sequence was very instrumental in the successful prediction of the 1975 M=7.3 Haicheng, China earthquake (Scholz, 1977). In other cases increased activity may be weaker or more spread in time, yet the events may be causally related to the mainshock and so should also be considered as foreshocks (Kanamori, 1981).

Jones and Molnar (1976), defining foreshocks as events occurring within 100km and 40 days of the mainshock, found that 44% of all large $M \geq 7$ shallow earthquakes in the world from 1950 to 1973 were proceeded by foreshocks large enough to be teleseismically reported, and that 21% of the large earthquakes that occurred in China between 1900 and 1949 had foreshocks large enough to be noted in local records. They note that the percent of large events with foreshocks is probably much higher than 44%. Many events with locally reported foreshocks (i.e. the 1975 Haicheng, China earthquake) had no foreshocks recorded teleseismically.

Many examples of possible broad sense foreshock activity have been reported. Mogi (1969a) and Kelleher and Savino (1975) show that for great earthquakes, seismic activity prior to failure tends to cluster around the future epicenter. Other examples of tight clustering of activity around future epicenters are given by Engdahl and Kisslinger (1977), Ishida and Kanamori (1978), and Fuis and Lindh (1979). A tabulation of many events for which foreshocks have been reported can be found in Rikitake (1975b, 1979) and Kanamori (1981).

The unambiguous identification of foreshocks sequences, as in the case of the 1975 Haicheng earthquake, would provide an extremely useful short term predictor for many earthquakes. Unfortunately, the occurrence of a foreshock sequence is generally not obvious until after the mainshock has occurred. Laboratory studies and theoretical considerations suggest that the ratio of smaller fractures to larger fractures should decrease as effective differential stress in an area increases (Scholz et al., 1973). This has lead to the postulate that foreshock sequences are characterized by smaller b values in the Gutenberg-Richter relation (equation 1) than are aftershock sequences or the background seismicity. Decreases in b have been reported for several foreshock sequences (for example, Suyehiro,1966; Papazachos et al., 1967; Suyehiro and Sekiya, 1972; Scholz et al, 1973; Papazachos, 1975; Wyss and Lee, 1973; Hasagawa et al., 1975). A large number of events are

needed for the reliable determination of b. Unfortunately, the number of recorded events in most foreshock sequences is small. The usefulness of monitoring local b values as a function of time to detect foreshock sequences is debatable. Monitoring this would require much more sensitive seismic networks than presently exist in most regions. Also, it is not clear that equation 1 is applicable to local seismicity patterns.

On the basis of a statistical study, Yamashina (1981) suggests that a sequence of events occurring within a short time interval of each other is more likely to be a foreshock sequence if the magnitude difference between the largest and second largest event is small. In chapter 2 of this thesis, it is suggested that, on the average, foreshocks are characterized by a greater rate of production of aftershocks. The above effects, if present, are weak. Their usefulness in identifying foreshock sequences is doubtful.

Ishida and Kanamori (1980) found that waveforms of foreshocks of the 1971 San Fernando and 1952 Kern County earthquakes are characterized by a concentration of energy at higher frequencies relative to earlier events. They suggest that such effects, although subtle, may be useful for monitoring the stress state of selected regions.

1.6.2 Swarms

A swarm can be loosely defined as a sequence of earthquakes which are close to each other in space, time, and magnitude (i.e. a sequence of tightly clustered earthquakes similar to an aftershock sequence but lacking the mainshock). McNally (1977) found that clusters of small earthquakes occurred close to the future epicenters of several moderate size earthquakes in California two to ten years before the mainshock. Ohtake (1976) and Sekiya (1977) reported similar findings for some larger Japanese earthquakes. Evison (1977a,b,c) finds that for many earthquakes worldwide which are characterized by quiescence, the quiescent period was preceded by a characteristic swarm of minor activity. He suggests that identification of swarms which are followed by quiescence may be useful for earthquake prediction.

1.6.3 Bursts of seismicity

Three related seismicity patterns, precursory to the largest earthquakes in a region, are proposed by Keilis-Borok et al. (1980a,b,c). These three patterns, collectively termed "bursts of seismicity", consist of the abnormal clustering of earthquakes in time, energy and space before a major earthquake.

The simplest of the three patterns is pattern B, "bursts of aftershocks". Pattern B consists of a medium magnitude mainshock with an anomalous number of aftershocks

concentrated at the beginning of the aftershock sequence. Pattern S consists of a "swarm" of mainshocks where a "swarm" is defined as a group of moderate events concentrated in space and time and occurring during a time interval when the overall seismicity is not below average. The final pattern, Sigma, consists roughly of an increase in the cummulative seismic energy released to the 2/3 power in a sliding time window. Sigma is roughly proportional to a summation of the fracture areas of the earthquake sources (Keilis-Borok and Malinovskaya, 1964). Pattern Sigma is identified as a peak in this summation.

In algorithmic form the three patterns have been applied with some success to many regions. Pattern Sigma was first observed for 20 large earthquakes in the Central Asia, Eastern Mediteranean, Pamir and Assam regions (Keilis-Borok and Malinovskaya, 1964). Gasperini et al. (1978) found that an overwhelming majority of strong earthquakes in Italy were preceded by anomalous "swarms" of weak events within 6 years of the mainshock. Keilis-Borok and Rotvain (1979) show that pattern S precedes strong ($M \geq 7.2$) earthquakes in Anatolia and surrounding regions. Sauber and Talwani (1980) apply pattern S to small earthquakes in the Lake Jocassee, South Carolina region. They show that the majority of $M \geq 2.0$ earthquakes are preceded by the pattern; a precursory time of 10 days was used. Keilis-Borok et al. (1980b) apply patterns B, S, and Sigma to retrospective prediction of strong ($M \geq 6.5$) earthquakes in Southern California. The

patterns were found to precede 5 out of 6 strong earthquakes in the region. Patterns Sigma and B were found to precede strong earthquakes in New Zealand; pattern B was reported for Italy. A precursory interval of 3 years was used in the Southern California, and New Zealand studies. Keilis-Borok et al. (1980a) found that 18 of 23 strong earthquakes, in five regions worldwide, were preceded by pattern B with a precursory time of 3 years. Keilis-Borok et al. (1980c) apply patterns Sigma and S to strong ($M \geq 8$) earthquakes in Tibet and the Himalayas.

In the above studies many parameters used in the definitions of the bursts of seismicity patterns were determined a posteriori. Some examples of free parameters which may or may not have been chosen a posteriori include: minimum magnitude of a strong earthquake in a region; precursory time interval; time windows for defining anomalous activity; boundaries of the region; definition of spatial size for the swarms of pattern S; minimum thresholds above which activity is considered anomalous; minimum magnitude considered for each pattern; and definition of aftershocks. The large number of free parameters involved constitutes the largest single difficulty in the evaluation of the significance of the above studies (Keilis-Borok et al., 1980a). When attempting to construct a hypothesis after the fact, the danger of self deception is considerable. By constructing a hypothesis with enough free parameters it may always be possible to fit the hypothesis to a known data set

even if the hypothesis is incorrect.

Bursts of seismicity patterns are regional in nature, unlike foreshocks, quiescence, and the swarms of McNally and Evison. As such, they do not indicate the location of the future earthquake. In most cases the patterns take place at some distance (hundreds of kilometers) from the mainshocks which they precede. Accordingly, the occurrence of patterns B, S and Sigma may be interpreted as indicating a significant increase in the probability of strong earthquake somewhere in the region within the next several years (Keilis-Borok et al., 1980b).

1.7 Quiescence and Activation

Many observed seismicity patterns can be interpreted as superpositions of quiescence and activation. The pattern of a precursory swarms followed by quiescence in the epicentral region suggested by Evison (1977a,1977b) is an example.

Quiescence in the epicentral region accompanied by increased activity in surrounding regions has been termed the "doughnut pattern". Mogi (1969a) reported this pattern for several large earthquakes in Japan. Yamashina and Inoue (1979) report a similar pattern before the magnitude 6.1 1978 Shimane earthquake.

An algorithm using a pattern similar to the doughnut pattern is given by Keilis-Borok and Rotvain (1979). This premonitory pattern is termed A-Q (activation-quiescence).

Pattern A-Q is defined as a combination of activation and quiescence on opposite sides of an earthquake prone site which is in a state of quiescence. The pattern presumes that the scheme of lineaments and sites where strong earthquakes are possible are known for the region of interest. Examples of regional lineaments and the determination of sites where strong earthquakes are possible are given by Gelfand et al. (1976) for California; Gelfand et al. (1974) for several regions in Central Asia, and Southeastern Europe; and by Gorshkov et al. (1979) for Italy.

Keilis-Borok and Rotvain (1979) show that for Anatolia and adjacent regions seven of nine $M \geq 7.2$ earthquakes which occurred from 1900 to 1976 were preceded by pattern A-Q with an average precursory time of 13 years. There were, however, many false alarms. Joint use of this pattern and pattern S ("swarms") gave better results (fewer false alarms) with a shorter average precursory time of 6 years.

Keilis-Borok and Rotvain (1978) suggest that pattern A-Q may be useful in the determination of soon-to-break seismic gaps.

Talwani (1979) gives an empirial earthquake prediction model for small earthquakes ($2.0 \leq M \leq 2.6$) in the Lake Jocassee area. He finds that premonitory seismicity is characterized by two distinct phases. The first phase, α phase, is characterized by a slow or no increase in overall seismicity and by quiescence in the epicentral area. The second phase, β phase, is characterized by an increase in

seismicity with a clustering of events around a "target" area defining the site of the future earthquake. The β phase is similar to the Doughnut pattern defined above. The mainshock occurred after a period of quiescence in the β phase. A similar pattern was observed before several large earthquakes in the Middle America region (Ohtake et al., 1977). The similarity between the micro-seismicity before small events at Lake Jocassee and reported precursory patterns for large events is remarkable.

1.8 Migration

Many investigators have noted an almost linear progression in time of large earthquakes along various seismic zones. An apparent northeasterly migration of large earthquakes through Northeastern China at a rate of 110km/yr was instrumental in the successful prediction of the 1975 M=7.3 Haicheng, China earthquake (Scholz, 1977). Other examples are numerous. Several authors have noted east to west migration at a rate of greater than 50km/yr along the North Anatolian fault of Northern Turkey (Mogi, 1968c; Kelleher, 1970; Toksöz et al., 1979). Toksöz et al. (1979) report a slower (≤10km/yr) easterly migration for the same zone. Mogi (1968c) observes a southeasterly migration pattern towards the 1933 Sanriku, Japan earthquake at a rate of 150km/yr. After the occurrence of the Sanriku earthquake the migration pattern changed direction towards the

southwest and continued at about the same rate away from the mainshock. Mogi (1968b,c) suggests a worldwide pattern of migration for large earthquakes ($M \geq 8.2$) in three branches between 1900 and 1964. In the Aleutains the apparent migration of two sequences of large earthquakes ($M \geq 7.7$) at a rate of around 100km/yr are noted by Kelleher (1970). Wood and Allen (1973) observe an apparent migration of earthquakes above magnitude 5 in Central California at a rate of 3.3km/yr. Bufe et al. (1974) observe an apparent migration of micro-earthquakes to greater depths before moderate earthquakes in Central California. Engdahl and Kisslinger (1977) find that the foci of foreshocks precursory to a magnitude 5 earthquake in the Central Aleutians gradually migrated towards the future mainshock epicenter. Using spectral analysis techniques Delsemme and Smith (1979) observe an apparent migration of large earthquakes ($M \geq 7.7$) in South America at a rate of 95km/yr from south to north

The existence of a migration pattern is difficult to prove. Migration patterns are usually based solely on visual observations and are therefore not amenable to statistical tests. An example of a quantitative statistical treatment of earthquake space time correlations is given by Kagan and Knopoff (1976, 1978). Using the second-order moment ("the number of events in different magnitude ranges that occur in a certain time-distance interval from a given shock") they demonstrate that a weak but statistically significant

migration pattern exists for large shallow earthquakes with migration velocities in the range of 300 to 2000km/yr.

1.9 Mainshock magnitude and precursory time

An apparent exponential dependency of precursor time on mainshock magnitude for some types of seismic precursors has been noted by several authors. Keilis-Borok and Malinovskaya (1964) find that the time difference between anomalous peaks in pattern Sigma and the subsequent mainshock increased with mainshock magnitude. A logarithmic dependence of precursor time on mainshock magnitude for several different types of long term precursory phenomena to moderate and strong earthquakes is reported by Scholz et al. (1973), Whitcomb et al. (1973), Evison (1977a,b), and Kristy and Simpson (1980). Talwani (1979) shows a logarithmic dependence of precursory time and mainshock magnitude for small ($2.0 \leq M \geq 2.6$) earthquakes in the Lake Jocassee region.

Several authors have reported apparent relationships between the maximum magnitude observed in the precursory seismicity and the magnitude of the mainshock. Papazachos (1974, 1975) and Wu et al. (1976) suggest that there is a direct relationship between the magnitude of the mainshock and that of the largest foreshock. However, Jones and Molnar (1979) find no such relationship. Evison (1977a,b) finds a linear relationship between the maximum magnitude of a precursory swarm and the magnitude of the mainshock. Kristy

and Simpson (1980) find a similar result for bursts of activity preceding large earthquakes in Central Asia.

Rikitake (1975b, 1979) shows that reported earthquake precursors can be grouped into separate categories on the basis of precursory time-mainshock magnitude dependence. Precursors of the first kind are characterize by a direct dependency of precursor time on the magnitude of future mainshock. This category includes most types of anomalous seismicity patterns with the exception of foreshocks. For short term precursors, including obvious foreshocks, no systematic correlation between precursor time and mainshock magnitude was found. A single exponential relationship between precursory time and mainshock magnitude seems to exist for all precursors of the first kind, although there is significant scatter in the data (Rikitake, 1975b, 1979). Other factors such as the regional strain rate may be important.

1.10 Criticism of previous works

The previous discussion has given examples of reported seismicity patterns. Most of the patterns are based solely on visual observations of epicenter maps or space-time diagrams of often poorly located events, as such the patterns are not well defined. A common procedure followed in looking for the patterns is to simply look at the seismicity in the immediate space-time vicinity of a large

event and to note any possible irregularities which seem to precede the large event. Different investigators have reported different and sometimes contradictory patterns for the same event. Little statistical significance can be given to patterns developed from observations on single events. Examples of more systematic studies are given by Keilis-Borok et al. (1980a,b,c), Keilis-Borok and Rotvain (1979), Sauber and Talwani (1980), Talwani (1979) and Kanamori (1981). Such systematic studies assume that there are systematic precursory patterns which do not vary substantially from event to event and region to region. This assumption may not be true.

What is needed in precursory studies is a quantitative definition for the patterns. For example, in the precursory patterns swarms (for example, Evison 1977a) and quiescence (for example, Mogi 1979) there is no quantitative definition of what a constitutes a precursory swarms or precursory quiescence. No studies are made to test for the occurrences of false alarms (i.e. the occurrences of "anomalous" swarms and periods of quiescence which are not followed by large events). Such studies need to be made before any statistical significance can be assigned to many of the reported patterns. Also needed is some physical model which explains the observed patterns.

In spite of the previous objections to the statistical significance of some observed precursory patterns, it should be noted that precursory patterns (in particular precursory

quiescence and precursory swarms) have been studied extensively by many independent investigators. Occurrences of precursory patterns are well documented, any model of sequences of earthquakes which fails to explain the observations suffers a severe deficiency.

1.11 On models of earthquake sequences

All observed precursory seismicity patterns even quiescence can be thought of as manifestations of clustering of earthquake sequences. Little is known about the physical nature of this clustering.

1.11.1 Stochastic models of clustering

In the literature many stochastic models for earthquake occurrence have been proposed which result in clustering of earthquakes in time. Such models are usually used to explain the statisics of aftershock sequences. A review of several such models can be found in Utsu (1972) and Vere-Jones (1975). An example is a time-dependent Poisson process in which the mean number of events per unit time is a function of time.

Another class of stochastic models with inherent clustering are the branching process models. Branching models start with one or a series of primary "events". These primary events generate secondary events with a probability that depends on the elapsed time from the primary event.

Secondary events generate subsequent events and so on. Branching models have been applied to models of crack expansion (Vere-Jones, 1976) as well as to statistical models of earthquake sequences (for example, Lomnitz, 1974; Kagan and Knopoff, 1981). The model of Kagan and Knopoff (1981) displays several interesting features and will be discussed below.

In the Kagan-Knopoff model the probability that one event gives "birth" to another in the time interval dt is given by

$$
\begin{aligned}
p(t)dt &= 0, && t < t_0 \\
p(t)dt &= (1-\alpha)\beta t_0^{\beta} t^{-(1+\beta)} dt, && t \geq t_0
\end{aligned}
\qquad 6
$$

The singularity at time zero is eliminated by assuming that a parent earthquake can not produce other shocks in time t_0 where t_0 is the assumed rupture time of an elementary earthquake. They suggest that a value of β equal to .5 gives results statistically similar to shallow earthquake sequences.

Using plausible assumptions for what would constitute an observed earthquake in such a model (ie the superposition of the effects of many elementary shocks which are spaced close in time), they show, through the statistical analysis of earthquake sequences generated by the model, that several time-magnitude properties similar to those observed in real catalogues are reproduced. These include the

frequency-magnitude law and the Omori law of the rate of aftershock production.

If such a model of earthquake occurrence is correct then present shocks can be thought of as a being the "children" of all previous earthquakes, or in a sense any earthquake is an "aftershock" of some earthquake occurring before it. All or most earthquakes will be interdependent. This presents a conceptual problem in dividing earthquakes into foreshock - mainshock - aftershock sequences if a purely statistical definition is assumed in which mainshocks are defined simply as those events which are statistically independent of other events and aftershocks and foreshocks are defined as statistically dependent events. This point will be discussed later. There is a related problem. We have only a short catalogue. In discussing statistical properties of the catalogues we are limited to time scales that are much less than the length of the catalogue, except for speculative purposes.

1.11.1.1 Deficiencies of stochastic models

A stochastic model of earthquake occurrence on the short data sets available can not be demonstrated to be correct. All that can be shown is that the model agrees with some observed properties of the sequences which are also extracted from short data sets. It should be remembered that the statisical properties of earthquake sequences are all based on a short data set. The short data set does not allow

for the description of longer term properties of the sequence. Omori's law postulates one such long term behavior. It should be remembered that Omori's law is based on a relatively short data set, there is no guarantee that it is valid for very long times. There is great ambiguity in what constitutes an aftershock; this ambiguity increases with the time from the mainshock. Any theory which rests on the vague notion of what is or is not a long term aftershock rests on a very weak foundation.

Stochastic models of earthquake occurrence ignore one very important aspect of the earthquake problem, namely the mechanical properties of the system. Little physical input is used to constrain the free parameters inherent to such models. There is no physical input in such models which accounts for the causes of seismicity (the energy source) or which takes into account the physical properties of the fault zone. Nothing is said about the fundamental cause of seismicity, namely the release of strain which accumulates due to plate motions. Stochastic models of earthquake sequences do not, nor can they, explain the observed precursory patterns. What is needed is some model which also incorporates the physical properties of the failure process and the mechanical properties of fault zones.

1.11.2 On the physical basis of clustering

Several qualitative physical explanations for earthquake clustering have been proposed. These include: inhomogeneities in the material strengths along a fault zone and non-linear dynamic friction on the fault zone.

1.11.2.1 Non-linear dynamic friction

Using a simplified two-dimensional model of the expansion of a crack Barenblatt et al. (1981) show that, with an appropriate assumption for the functional form of the dynamic friction as a function of slip rate and the modulus of cohesion as a function of the rate of expansion of the crack, expansion of the crack is realized by alternating slow and fast phases. The number of alternations becomes large in the presence of local maxima in the shear stress or local minima in the friction along the fault. They suggest that the effect may be a possible explanation for the occurrence of premonitory clusters at large distances relative to subsequent strong earthquakes.

The functional form of the dynamic friction as a function of slip rate has the following characteristics: for small slip rate the dynamic friction increases with increasing slip rate; for intermediate values of slip rate the dynamic friction decreases with increasing slip rate to some minimum; after the minimum is achieved dynamic friction increases steadily with increasing slip rate. The modulus of cohesion is assumed to increase with rate of crack

expansion. There is little data supporting the assumed functional form of the dynamic friction as a function of slip rate.

1.11.2.2 The barrier or asperity model

Irregularities in the physical or mechanical properties along a fault zone provide another possible explanation for clustering. Regions of high strength along a fault zone are usually termed asperities or equivalently barriers. The concept of asperities has been used to qualitatively explain many features of seismicity. These include, for example: complex events (Das and Aki, 1977; Aki, 1979; Kanamori and Stewart, 1978; Lay and Kanamori, 1980; and others); non-uniform seismicity along fault zones (Wesson and Ellsworth, 1973); the dependence of the frequency of foreshocks on the time to the mainshock (Jones and Molnar, 1979); and seismic clustering (Ishida and Kanamori, 1978, 1980).

Kanamori (1981) shows that a simple asperity model can explain many features of the various observed precursory seismicity patterns. The following will give a discussion of this model since it has great relevance to this work.

The model assumes that a fault is composed of a series of subfaults with varying strengths. The strength of a subfault is defined as the maximum loading stress it can acommodate before it fails. A region which is characterized by the occurrence of large earthquakes is a zone which is

characterized by an increase in the strength of subfaults. Accordingly, an asperity is introduced as a region of high strength of subfaults relative to the "normal" distribution of strengths. As a simple example Kanamori (1981) considers the case with one asperity. The strengths on the subfaults are assumed to follow a gaussian distribution with mean S and standard deviation α. The strengths of segments on the asperity are assumed to follow a similar distribution with larger mean (S_0) and with different standard deviation (α_0). The model further assumes that a section of the fault fails when the stress on that section exceeds the strength and once a section fails the loading stress it accommodated is held uniformly by the remaining unfailed sections. Once a subfault fails the stress there drops to zero and no healing occurs. In such a model the rate of stress increase on unbroken sections accelerates as the number of broken sections increases.

Consider qualitatively what happens when the loading stress increases linearly in time in such a situation. When the tectonic loading stress is low compared to S only a small numbers of subfaults will fail as a scattered activity. As the loading stress increases approaching S the process accelerates until most subfaults outside the asperity are broken. The increased activity at this stage could result in swarm-like activity since the loading stress will exceed S rapidly.

After this stage quiescence could result since the sections of the fault which remain are relatively strong. The stress will be concentrated on the asperity, the area surrounding the asperity will be essentially decoupled. This may also result in loading of adjacent fault zones which could produce a doughnut pattern in the seismicity. In the final stage as the loading stress approaches the strength of the asperity S_0 the asperity itself will begin to break in an accelerated process possibly resulting in a concentrated swarm which terminates with the mainshock. Kanamori (1981) gives a simple numerical simulation which reproduces the above behavior. The time length of the periods of quiescence and of the precursory swarm which terminates with the large event are controlled by the parameters S_0/S and α_0. As S_0/S decreases the length of the quiescence decreases. As α_0 increases the length of the precursory swarm which terminates with the large event increases. If α_0 is too small no such precursory swarm appears.

Such a process, although simplistic, gives a possible qualitative explanation of swarms, quiescence, and activation before large earthquakes. It does not explain the large distances and times which are sometimes attributed to precursors.

Furthermore, the model suffers from too many free parameters. In such a qualitative picture it might be possible to "model" any sort of activity by varying the number and characteristics of the asperities. Nothing is

said about the nature of the interaction between different fault zones. No healing mechanism is present. Such a model can not explain the apparent recurrence of large earthquakes at nearly the same location without some sort of healing mechanism for the asperities. However, the model is perhaps useful as the simplest possible model which could explain the complexity of observed seismicity patterns.

1.12 Concluding remarks

Anomalous seismicity has been observed before a large number of earthquakes in nearly all magnitude ranges. Most patterns are based solely on visual observations of epicenter maps or space-time diagrams. Such visual patterns are only qualitatively defined. Some examples of attempts at quantitative descriptions of seismicity patterns are given by Keilis-Borok et al. (1980a,b,c), Keilis-Borok and Rotvain (1979), Gasperini et al. (1978), Sauber and Talwani (1980), and Talwani (1979). Different investigators have reported different and sometimes contradictory patterns for the same event. Quiescence seems to be the most commonly observed pattern.

In a global survey of seismicity patterns before major earthquakes along subduction zones Kanamori (1981) finds significant regional variations in the pattern of which areas have foreshocks, quiescence, and premonitory swarms.

Although there are significant variations in the premonitory patterns for individual events, on the average different magnitude ranges show remarkable similarities if appropriate time and space scaling is taken into account. This may be another manifestation of the self similarity of the earthquake process proposed by Kagan and Knopoff (1980). If so, the study of micro-seismicity preceding relatively numerous small events may be useful in determining the nature of seismicity before large earthquakes.

Many patterns may not be sufficiently obvious to be observed before the occurrence of the mainshock. Also, it is not known how many false alarms would result from the use of the various patterns (for example, there have been no studies on the numbers of anomalous swarms and periods of quiescence which were not followed by large earthquakes). This along with variations in patterns from event to event limits the usefulness of seismicity patterns alone in earthquake prediction. However, with other constraints such as estimates of recurrence intervals and locations of possible strong earthquakes, seismicity patterns may be useful in forecasting the times, places, and perhaps magnitudes of future earthquakes at least on a long-term basis.

2. A scaling law for the occurrence of aftershocks in Southern California

2.1 On self similarity

The concept of stochastic self-similarity is a familiar and important concept in the field of turbulent flow. The self-similarity of turbulent flow implies that the "picture" of turbulence shows no characteristic scale. The picture of flow will look statisically similar at any magnification between an outer scale, determined by the driving force, and an inner scale, where viscosity becomes important. Recent works of Kagan and Knopoff (1976, 1977, 1978, 1980a, 1980b) suggest that seismic process shows similar stochastic self-similarity, (i.e. there seems to be an absence of any particular scale connected with time - distance - magnitude patterns of earthquake occurrence; epicentre maps have the same appearance regardless of scale). Kagan and Knopoff (1978, 1980b) report that this self-similarity holds over magnitude ranges extending from M=1.5 to the largest earthquakes. Insufficient data is available to test the hypothesis for M<1.5.

Andrews (1980) shows that the Gutenberg Richter frequency magnitude relation of earthquake occurrence results by assuming self-similarity alone. Furthermore, self-similarity manifests itself in the form of known power law distributions of many features of seismicity. These

include: the power law distribution of seismic energies and Omori's law for the rate of occurrence of aftershocks and foreshocks of shallow events (Kagan and Knopoff, 1978); the power-law dependence of the energies of foreshocks and aftershocks (Kagan and Knopoff, 1978); and the inverse power-law dependence of the spatial moment on the separation between two foci (Kagan and Knopoff, 1980a).

The apparent self-similarity of the failure process has great relevance to the identification of anomalous seismicity patterns. In a given region large events are rare. The time spans of earthquake data sets are usually short compared to the return times of the largest events in a region. This means that the statistical basis of any phenomenon associated with large magnitude earthquakes will at best be very weak. Self-similarity implies that the study of anomalous seismicity in the occurrence of small events before relatively numerous moderate events may be useful in the identification of anomalous seismicity in the sequence of the moderate events themselves, provided that appropriate time and space scaling is taken into account.

This chapter explores an aspect of the self-similarity of the earthquake process, a scaling law for the occurrence of aftershocks in Southern California. The research which forms the basis of this chapter was carried out at the Seismological Laboratory, California Institute of Technology. The chapter is based on the paper "A scaling law for the occurrence of aftershocks in Southern California" by

R. Lamoreaux, V. I. Keilis-Borok, and K. H. Hutton which has been submitted to *Physics of the Earth and Planetary Interiors*.

2.2 Introduction

The occurrence of aftershocks has been intensively studied in many branches of seismology. These include: statistical models of the earthquake source (Utsu, 1972; Vere-Jones, 1976; Kagan and Knopoff, 1976); mechanical models of the earthquake source (Aki, 1979; Kanamori, 1981; Barenblatt et al., 1981); and estimation of seismic risk and earthquake prediction. Some of the premonitory seismicity patterns discussed in Chapter 1 such as seismic gaps (Sykes, 1971) and "bursts of aftershocks" (Keilis-Borok et al., 1980a, 1980b) are directly based on the observed pattern of aftershock occurrence.

The above mentioned studies require a knowledge of statistical properties of aftershock sequences. Not many such properties are known. Utsu (1972) gives a review. One of the few comparatively well established regularities is the Omori Law

$$n(T(e)) = C/[T(e)+\beta]^{\alpha} \qquad 7$$

where n(T(e)) is the average number of aftershocks during the eth day after main shock, C, α and β are constants. The coefficient C obviously depends on the magnitude of the

mainshock and the magnitude range in which aftershocks are counted. It also depends on the location of the hypocenter and probably on the mechanism of the main shock. However, even when all the above factors are similar, n(T(e)) shows strong variations from event to event. Omori's law only has meaning as a rough average for many events. As such, it is valid for different formal defintions of aftershocks (Gardner and Knopoff, 1974; Vere-Jones, 1976; Keilis-Borok et al., 1971). The commonly accepted values for α and β are: $\alpha=1$ and $\beta=0$. The large number of factors which influence n(T(e)) presents a great difficulty in the estimation of C, α and β. There are probably too many parameters for the available data.

This chapter will test the hypothesis that the number of aftershocks does not depend on the magnitude Mm of the mainshocks if aftershocks are counted in a magnitude interval from (Mm-Δ) to Mm, where Δ is a constant. The total number of aftershocks over a period of the order of a year is considered. In terms of Omori's law, the hypothesis means that for a fixed Δ, C does not depend on Mm. This hypothesis, if correct, provides a self-similarity which imposes an additional constraint on the study of the physical nature of aftershocks. Also, the elimination of one parameter (Mm) reduces the number of observations necessary for statistical studies of aftershock sequences.

In the next section the above mentioned scaling law is tested for Southern California. The Southern California

catalogue 1932 through 1980 (Hileman et al., 1973; Friedman et al., 1976; Subsequent Preliminary Epicenter Listings) provided the data for the study. Finally the scaling law is used for the analysis of a long-term premonitory seismicity pattern termed "bursts of aftershocks" (Keilis-Borok et al., 1980a, 1980b). This pattern is also discussed in Chapters 1 and 4. Keilis-Borok and Prozoroff (1981) give a similar study for the strongest earthquakes worldwide ($M \geq 7$).

2.3 The scaling law

To test the hypothesis formulated in the introduction, the number of aftershocks for different mainshock magnitude Mm was counted for several values of Δ. The following simple definition for an aftershock was used: Consider two earthquakes with time sequence numbers i and j, i>j. The second earthquake is an aftershock of the first if the following conditions are satisfied: The distance between their epicenters is less than R(Mi); the time difference $ti-tj \leq T(Mi)$; and $Mj \leq Mi$. T(M) and R(M) are empirical functions. There is no physical basis for this definition. The functions T(M) and R(M) can be viewed as simple box car windows in which obvious "mainshock-aftershock" clustering is evident. The thresholds R(M) and T(M), assumed after Gardner and Knopoff (1974) are given in Table 2.1. They seem to give an estimation from above (i.e. for most mainshocks R(M) and T(M) are larger than necessary). An exception is

Windows for the identification of aftershocks (after Gardner and Knopoff, 1974).

M	R(km)	T(days)
2.5	19.5	6.0
3.0	22.5	11.5
3.5	26.0	22.0
4.0	30.0	42.0
4.5	35.0	83.0
5.0	40.0	155.0
5.5	47.0	290.0
6.0	54.0	510.0
6.5	61.0	790.0
7.0	70.0	915.0
7.5	81.0	960.0
8.0	94.0	985.0

Table 2.1 gives the empirical windows for the identification of aftershocks (after Gardner and Knopoff, 1974).

the Kern County earthquake, July 1952, M=7.2. For this earthquake R(M) and T(M) are greater than those values indicated in table 2.1. A map of mainshocks with magnitudes greater than or equal to 4.0 for the years 1932-1980 using the parameters of table 2.1 is given in figure 2.1.

The total number of aftershocks (bi) were considered. It follows that

$$b_i=\sum_{e=1}^{T(M)} n(T(e)) \qquad 8$$

where n(T(e)) is given by (1). The results of the computations are shown in tables 2.2 - 2.4. It is presumed that the Southern California catalogue is reasonably complete for $M \geq 3.5$ from the beginning (1932). This allowed the computation of bi for $\Delta=1$, $Mm \geq 4.5$ for mainshocks from 1932. For the years 1970 - 1980 earthquakes with magnitudes greater than or equal to 3.0 were considered and for 1977 - 1980 events with M greater than or equal to 2.5 were considered. These divisions are approximate.

The above places obvious constraints on the miminum mainshock magnitude which could be considered for the different values of Δ. The number of smaller magnitude mainshocks occurring after 1977 was insufficient to test the hypothesis for $\Delta=2.0$. To test the hypothesis for $\Delta=2.0$ the above constraint was relaxed; $M\geq 2.5$ from 1970 and $M\geq 2.0$ from 1977 were included. The results for $\Delta=2.0$ could be biased due to this.

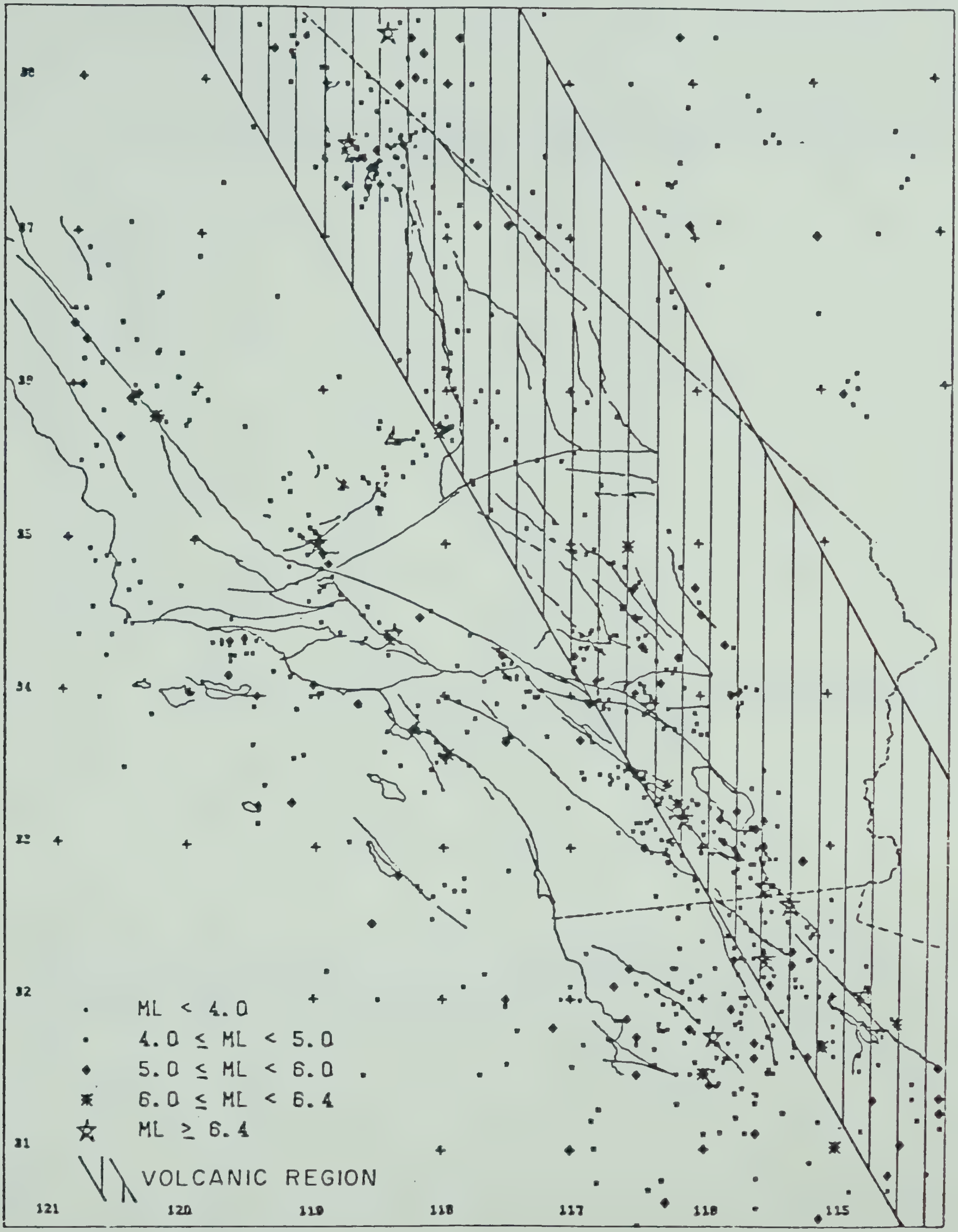


Figure 2.1 gives an epicentral map of mainshocks occurring in Southern California from 1932 through 1980 as determined by the aftershock criteria of table 2.1. The approximate contour of the volcanic zone after (Allen et al., 1965) is shaded.

Frequency distribution of numbers of aftershocks per mainshock, delta = 1.0

Number of aftershocks	4.0-4.4	4.5-4.9	5.0-5.4	5.5-5.9	6.0-6.4	>6.4
>15	-	-	-	-	-	-
15	1	-	-	-	-	-
14	-	-	-	-	-	-
13	-	-	-	-	-	-
12	-	1	-	-	-	-
11	-	-	-	-	-	-
10	-	-	1	-	-	-
9	-	2	-	-	-	-
8	-	-	-	-	-	-
7	1	-	-	1	-	-
6	-	3	-	-	-	-
5	1	6	1	1	-	-
4	4	4	2	-	-	-
3	3	10	1	3	2	-
2	12	13	6	-	2	-
1	15	39	19	6	3	4
0	31	96	24	14	2	3

Mainshock Magnitude

Table 2.2 is a tabulation of the frequency of occurrence for the number of aftershocks per mainshock with $\Delta = 1$. The mainshock magnitude range 4.0 to 4.4 uses data after 1970, other mainshock magnitude ranges use data from 1932.

Frequency distribution of numbers of aftershocks
per mainshock, delta = 1.5

Number of aftershocks	4.0-4.4	4.5-4.9	5.0-5.4	5.5-5.9	6.0-6.4	>6.4
>20	5	1	1	1	1	-
20	-	-	-	-	-	-
19	-	-	-	-	-	-
18	-	-	-	-	-	1
17	-	-	1	-	-	-
16	-	-	1	1	-	-
15	-	-	-	1	-	-
14	-	-	-	-	-	-
13	-	-	-	-	-	-
12	1	-	2	-	-	-
11	-	-	-	-	-	-
10	-	-	-	-	-	1
9	-	-	-	-	-	1
8	-	-	1	1	1	-
7	2	1	1	1	1	-
6	2	-	-	-	-	1
5	7	1	2	3	1	1
4	10	1	3	-	3	1
3	6	2	7	3	-	-
2	9	4	7	3	2	-
1	12	5	11	4	-	1
0	16	9	15	9	-	3

Mainshock Magnitude

Table 2.3 is a tabulation of the frequency of occurrence for the number of aftershocks per mainshock with $\Delta = 1.5$. The mainshock magnitude range 4.0 to 4.4 uses data after 1977, 4.5 to 4.9 data after 1970, other mainshock magnitude ranges use data from 1932.

Frequency distribution of numbers of aftershocks
per mainshock, delta = 2.0

Number of aftershocks	4.0-4.4	4.5-4.9	5.0-5.4	5.5-5.9	6.0-6.4	>6.4
>25	4	1	2	3	1	-
24-25	1	-	-	-	-	-
22-23	-	-	-	-	-	-
20-21	1	-	-	1	-	1
18-19	-	-	1	-	-	-
16-17	-	1	1	-	2	-
15	1	-	1	-	-	-
14	-	1	-	1	1	1
13	1	-	-	2	-	-
12	-	-	2	-	2	-
11	2	-	-	3	-	-
10	-	1	-	2	2	1
9	-	1	-	-	-	1
8	1	1	-	-	-	-
7	2	1	-	-	-	-
6	2	-	-	1	-	-
5	1	3	2	1	-	1
4	-	1	-	2	-	1
3	-	1	-	-	-	-
2	-	3	-	1	-	-
1	-	5	-	2	-	-
0	2	5	1	7	-	2

Mainshock Magnitude

Table 2.4 is a tabulation of the frequency of occurrence for the number of aftershocks per mainshock with $\Delta = 2$. The mainshock magnitude range 4.0 to 4.4 uses data after 1977, 4.5 to 5.4 data after 1970, other mainshock magnitude ranges use data from 1932.

Tables 2.2, 2.3 and 2.4 show frequency distributions of b versus mainshock magnitude for $\Delta=1.0$, 1.5, and 2.0 respectively. Figures 2.2 -2.4 show the corresponding normalized cumulative frequency distributions b(q), where q is the number of mainshocks with b or less aftershocks. The hypothesis predicts that the frequencies tabulated in tables 2.2 - 2.4 should be independent of mainshock magnitude interval. To test this each table was divided into magnitude intervals such that each interval contained similar numbers of events. The frequencies in the magnitude intervals were then compared by the Chi-squared criteria. The results are shown in table 2.5. The hypothesis that the frequencies are the same cannot be rejected at the .28 level of significance for the worst case ($\Delta=1.0$).

Tables 2.2 - 2.5 and figures 2.2 - 2.4 seem to confirm our hypothesis at least for $\Delta\leq1.5$ and Mm>4.0. The catalogue is insufficiently complete to make conclusions for smaller Mm or larger Δ. For $\Delta=2.0$ the results are encouraging, although for later years the minimum aftershock magnitude used in this case is probably not justified by catalogue completeness. The cumulative frequency distributions for different Δ are juxtaposed on figure 2.5.

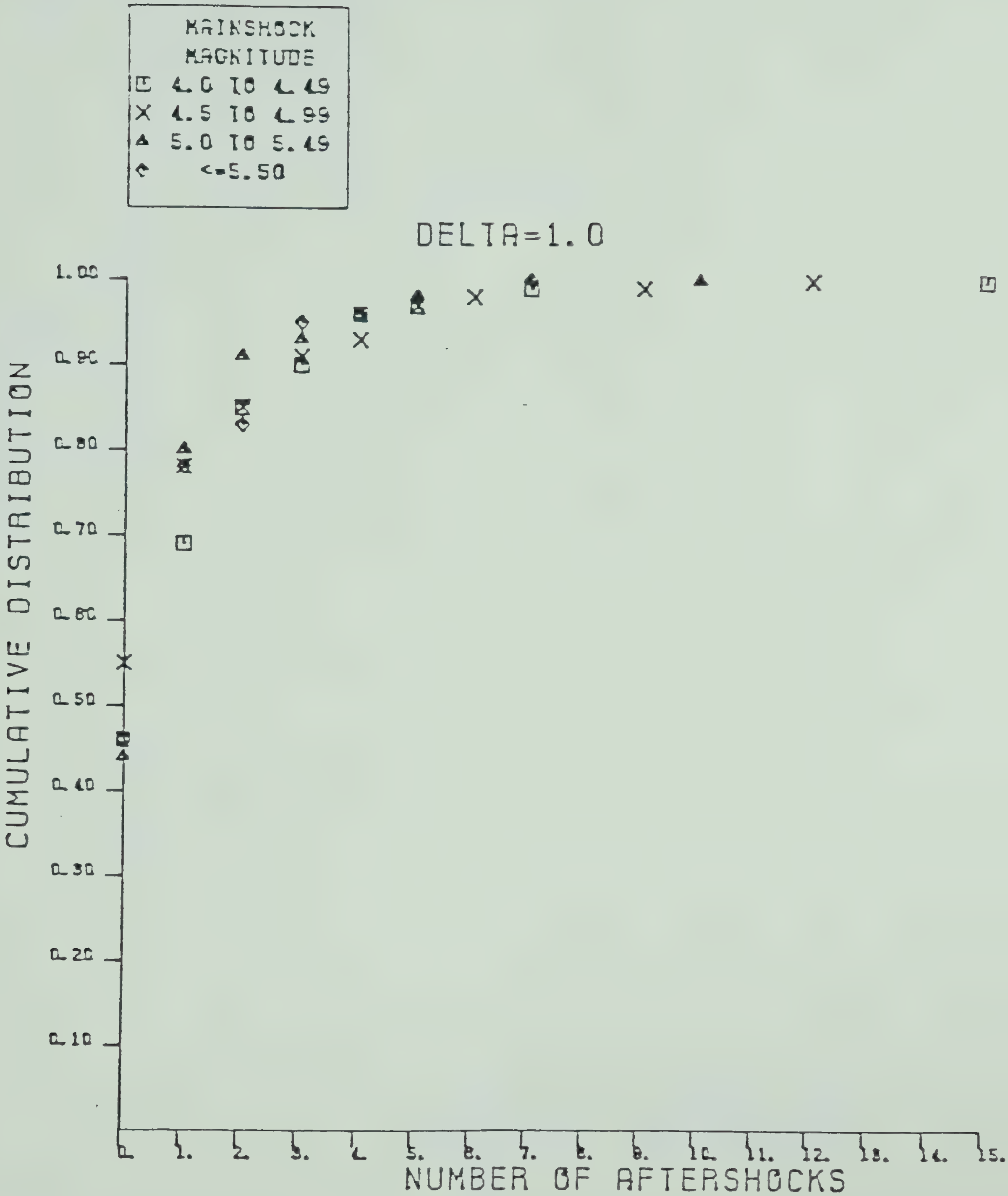


Figure 2.2 gives the normalized cumulative frequency distribution b(q) $\Delta = 1$ for various magnitude ranges.

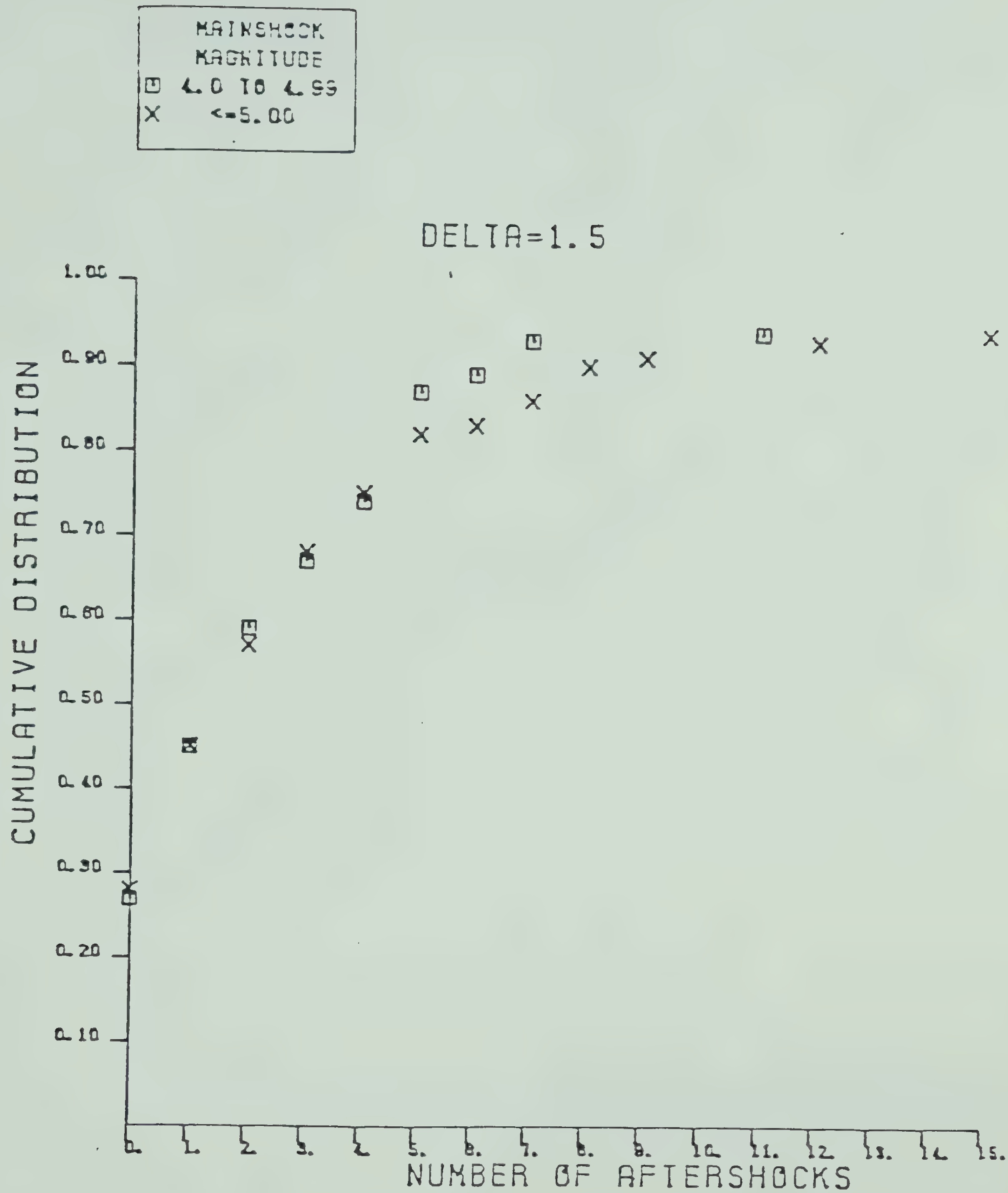


Figure 2.3 gives the normalized cumulative frequency distribution b(q) for Δ = 1.5.

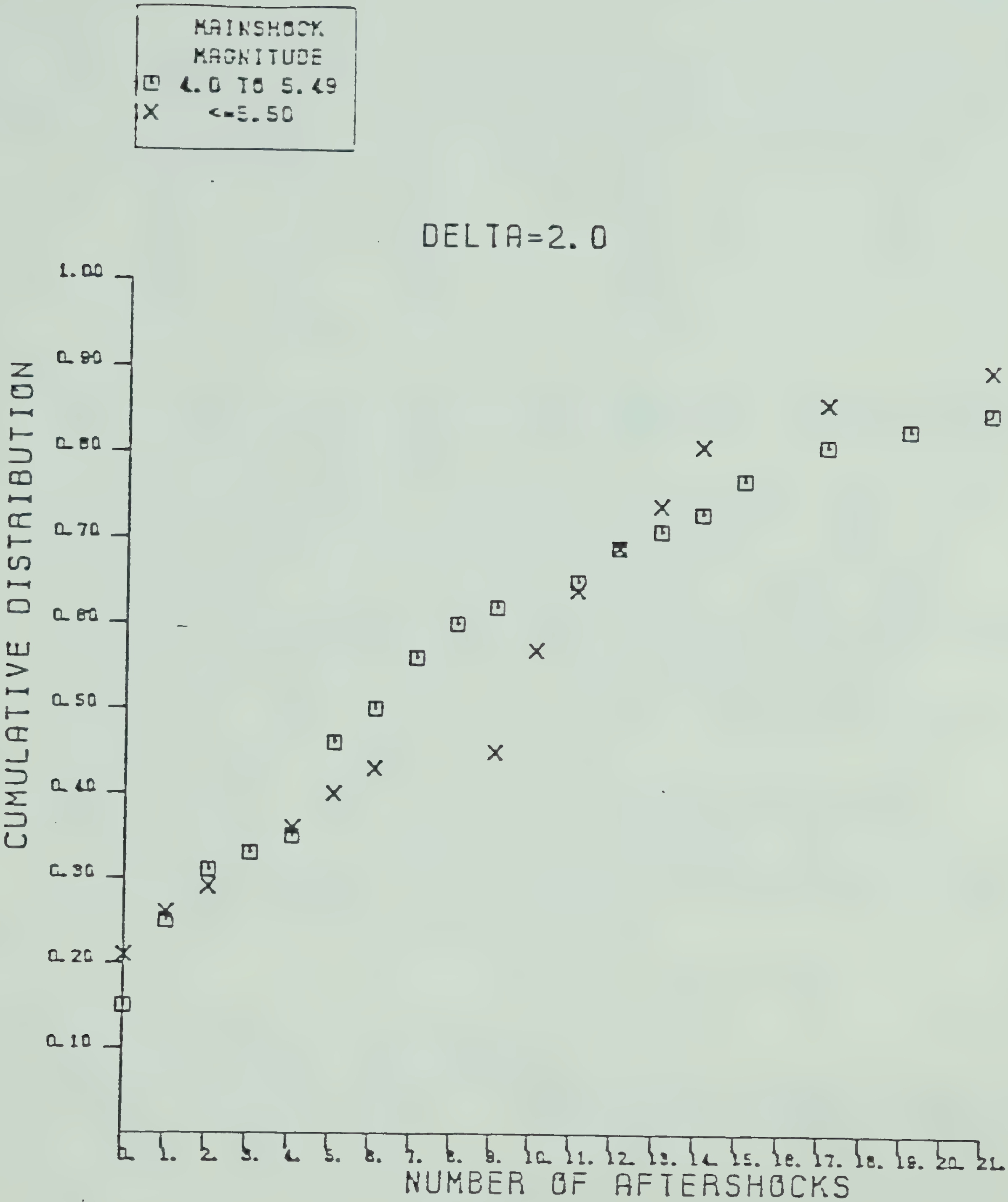


Figure 2.4 gives the normalized cumulative frequency distribution b(q) for $\Delta = 2.0$.

Chi-square comparisons of the frequency distributions

Key	delta	x	q	P
(A)	1.0	2.55	2	.72
(B)	1.5	.30	3	.04
(C)	2.0	2.77	3	.57
(D)	1.5	7.26	3	.94
(E)	1.0	7.21	1	.99

(A) gives a comparison between the magnitude range 4.0-4.9 for the years 1970-1980 and the magnitude range >4.9 for for the years 1932-1980.
(B) compares the magnitude ranges 4.0-4.4 after 1977 plus 4.5-4.9 after 1970 to the ranges >4.9 after 1932.
(C) compares the magnitude range 4.0-4.4 after 1977 plus 4.5-5.4 after 1970 to the combined distribution for all magnitudes greater than 5.4 occurring from 1932.
(D) compares the combined distribution of the magnitude ranges 4.0-4.4 (1977-1980), 4.5-4.9 (1970-1980) and >4.9 (1932-1980) which occurred in the volcanic region outlined in figure 2.1 to the similar distribution for the non-volcanic region.
(E) compares the frequency distributions of earthquakes found to be foreshocks to the expected distribution for mainshocks. The number of aftershocks was counted for two days. Earthquakes with magnitudes greater than 4.4 were considered for all years. Those with magnitudes 4.0-4.4 were considered after 1970.

q gives the number of degrees of freedom
x gives the value of Chi-squared
P is the Probability that a random variable which follows the Chi-square distribution with n degrees of freedom is less than or equal to x.

Table 2.5 gives the Chi-square test of goodness of fit between various magnitude intervals from the frequency tables 2.2-2.4.

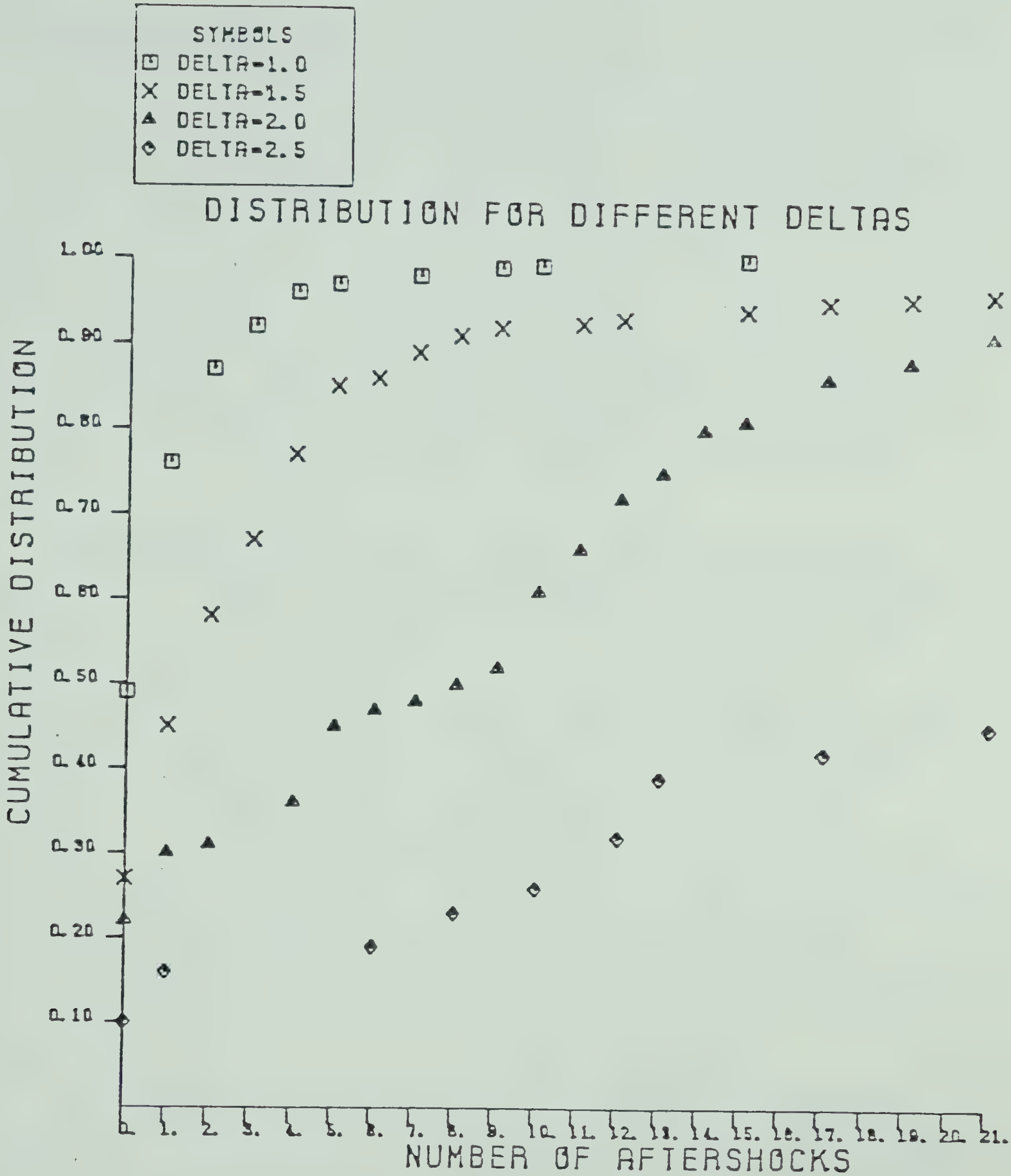


Figure 2.5 shows a comparison of the normalized cumulative frequency distributions for different values of Δ. The distributions are based on all mainshocks for which the catalogue was considered sufficiently complete (ie $M_m \geq m_o(t) + \Delta$, where $m_o(t)=3.5$ until 1970, 3.0 from 1970 to 1977 and 2.5 after 1977). The distributions for $\Delta=2.0$ and $\Delta=2.5$ are based on small numbers of points (64 and 31 respectively).

2.4 Regional variations

Examination of the catalogue reveals that the clustering of earthquakes is different in different parts of Southern California. For example, the area of Quaternary volcanism is known as "aftershock prone" (Allen et al., 1965). This area is characterized by more intensive clustering, i.e. clusters with larger numbers of events. Accordingly, b should be larger on the average in this region.

The approximate contour of the volcanic area (after Allen et al., 1965) is shown in figure 2.1. The cumulative frequency distribution of b for the volcanic and non-volcanic areas with $\Delta=1.5$ are given in figure 2.6. The differences between the two regions are evident from the figure. For $\Delta=1.5$ the frequency distribution for the volcanic zone is characterized by fewer mainshocks with zero aftershocks (23% compared to 33% for the non-volcanic region) and more mainshocks with greater than 20 aftershocks (7% compared to 0% in the non-volcanic region).

The frequency distributions for the two areas were compared by Chi-squared calculations. The results are shown in table 2.5. The hypothesis that the distributions are similar can be rejected at the .06 level of significance.

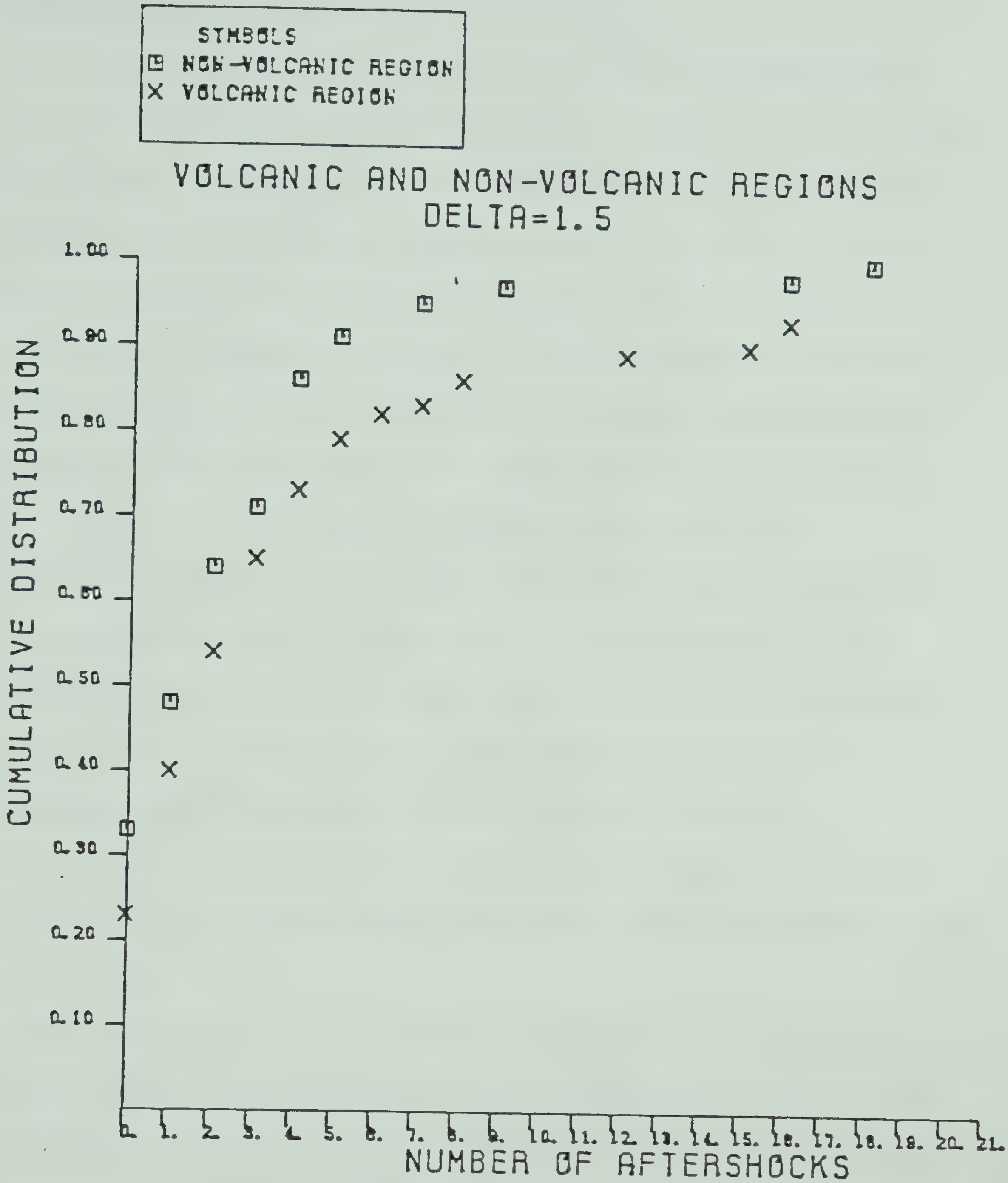


Figure 2.6 give a comparison of the normalized cumulative frequency distributions b(q) for the volcanic and the non-volcanic regions (Δ=1.5).

2.5 Premonitory pattern B

According to Keilis-Borok et al. (1980a, 1980b) the occurrence of a "burst of aftershocks" (pattern B) indicates a significant increase in the probability of a strong ($M \geq M_0$) earthquake in the same region within the next t_0 years. "Pattern B" consists of a mainshock in the medium magnitude range with an anomalous number of aftershocks concentrated at the beginning of the aftershock sequence. Specifically the pattern is a mainshock with magnitude Mi such that $M_2 \leq Mi \leq M_1$ and $bi(e) \geq \bar{B}$ where bi(e) is the number of aftershocks in the first e days following the mainshock and $\bar{B}$ is a threshold above which bi(e) is considered to be anomalous. The following values were selected for Southern California by retrospective juxtaposition of strong earthquakes and the values of bi: $M_0=6.5$; $M_1=M_0-.1$; $M_2=M_1-1.$; $t_0=3$ years; $\bar{B}=13$; and e=2days (Keilis-Borok et al., 1980b). Aftershocks were counted in the magnitude range from 3 to Mm.

The conclusion of the previous section allows us to choose $\bar{B}$ independently of Mm and juxtaposition with strong earthquakes, although not unambigiously. We define $\bar{B}(q)$ as the q% quantile of the empirical distribution of the values of bi (i.e. $\bar{B}(q)$ is the value of b which is greater than the bi values of q% of the mainshocks).

Completeness of the catalogue restricts the maximal value of Δ we can use. For Mm=5.5 we are restricted to $\Delta \leq 2.0$ for early years. We considered the distribution of bi(2

days) for Δ=2.0, Mm>5.5, for the years 1932 - 1970, Mm>5.0 for 1970 - 1977, and Mm>4.5 for 1977 - 1980 and obtained $\bar{B}(95) = 13$ and $\bar{B}(90) = 11$.

Mainshocks with M>5.0 and bi(2 days)>11 are juxtaposed with strong earthquakes in table 2.6. As expected the results are very similar to those obtained by Keilis-Borok et al. (1980b). Most events found to be anomalous previously are found to be anomalous here. The exceptions are listed in table 2.7. Seven of nine strong events are preceded by anomalous bi. There are two false alarms, one in 1935 and one in 1958. The false alarm in 1935 is generated by long distance aftershocks of the 1934 M=6.5 and M=7.1 strong events.

For smaller Δ the identification of anomalous events becomes more sensitive to the magnitude estimates, particularly to the mainshock magnitude. For example, Keilis-Borok et al. (1980b) identified the Oct. 4, 1978 Mm=5.8 earthquake as anomalous with bi(2 days)=35. For Δ=2.0 bi(2 days)=9 which is not identified as anomalous at the 90% quantile level. However, if the mainshock magnitude were over estimated by .1 two more aftershocks should be included and bi=11 which would be identified as anomalous. An increase in Δ would make possible magnitude errors less significant as well as increase the aftershock numbers. This would make the separation of anomalous events from non-anomalous events clearer. It is therefore suggested that as large a magnitude range as is possible should be used to

Mainshocks with anomalous bi (delta=2.0)
and strong earthquakes

Year	day	M	lat.	Long.	bi	*bi
1933	364	6.3	33.62	-117.97	44	103
1934	365	7.1	32.25	-115.50		
1935	119	5.0	31.75	-116.50	15	-
1935	251	5.0	32.90	-115.22	11	-
1940	140	6.7	32.73	-115.50		
1941	182	5.9	34.37	-119.58	11	18
1942	294	6.5	32.97	-116.00		
1947	205	5.5	34.02	-116.50	15	29
1948	339	6.5	33.93	-116.38		
1950	209	5.4	33.12	-115.57	12	-
1950	210	5.5	33.12	-115.57	13	13
1952	203	7.2	35.00	-119.02		
1954	316	6.3	31.50	-116.00	17	19
1955	351	5.4	33.00	-115.50	11	-
1956	40	6.8	31.75	-115.92		
1958	334	5.8	32.25	-115.75	11	11
1969	80	5.8	31.20	-114.20	44	49
1971	40	6.4	34.41	-118.40		
1979	1	5.0	33.94	-118.68	14	
1979	74	5.2	34.33	-116.44	14	
1979	288	6.6	32.61	-115.32		
1980	146	6.5	37.56	-118.79		

bi was counted for a 2-day interval after the mainshocks.
*bi indicates values of bi from Keilis-Borok et al., (1980b).

Table 2.6 lists mainshocks with anomalous bi for $\Delta = 2.0$ along with strong earthquakes $M_0 \geq 6.4$. Strong earthquakes are underlined.

Mainshocks with anomalous bi from Keilis-Borok et al. (1980b) not found to be anomalous for delta = 2.0

Year	day	M	lat.	Long.	bi	*bi
1937	84	6.0	33.41	-116.26	5	13
1946	74	6.3	35.73	-118.05	6	13
1947	100	6.2	34.98	-116.55	8	16
1949	122	5.9	34.02	-115.68	2	19
1954	78	6.2	33.28	-116.18	9	24
1968	100	6.4	33.19	-116.13	4	13
1978	277	5.8	37.51	-118.68	9	22

bi was counted for a 2-day interval after the mainshocks.
*bi indicates values of bi from Keilis-Borok et al., (1980b).

Table 2.7 list mainshocks with anomalous bi from Keilis-Borok et al. (1980b) which were not found to be anomalous for $\Delta = 2.0$ and $\bar{B}(q) = 90\%$.

identify anomalous bi.

2.6 Aftershocks of foreshocks

In the test of the scaling law we have eliminated events found to be foreshocks from our analysis. The definition of foreshock used here is similar to the previously mentioned definition for aftershocks with the exception that the condition Ma<Mm is reversed to Mf<Mm, where Mf is the magnitude of the foreshock. The same thresholds R(Mf) and T(Mf) were assumed.

The number of aftershocks bi(e) with Δ=1.0 was computed for each foreshock for which tm-tf>e. A value of 2 days was chosen for e. Figure 2.7 shows the corresponding normalized cumulative frequency distributions of bi(2 days) for foreshocks and mainshocks. The frequency distribution for foreshocks was compared to the expected frequency distribution generated by mainshocks using the Chi-squared criteria. The result is given in table 2.5. The hypothesis that the distribution for foreshocks is the same as that for mainshocks can be rejected at the .01 level of significance.

Figure 2.7 leaves the impression that on the average foreshocks are characterized by larger b. However, the total number of foreshocks, 27, is too small for definite conclusions. This difference deserves further attention.

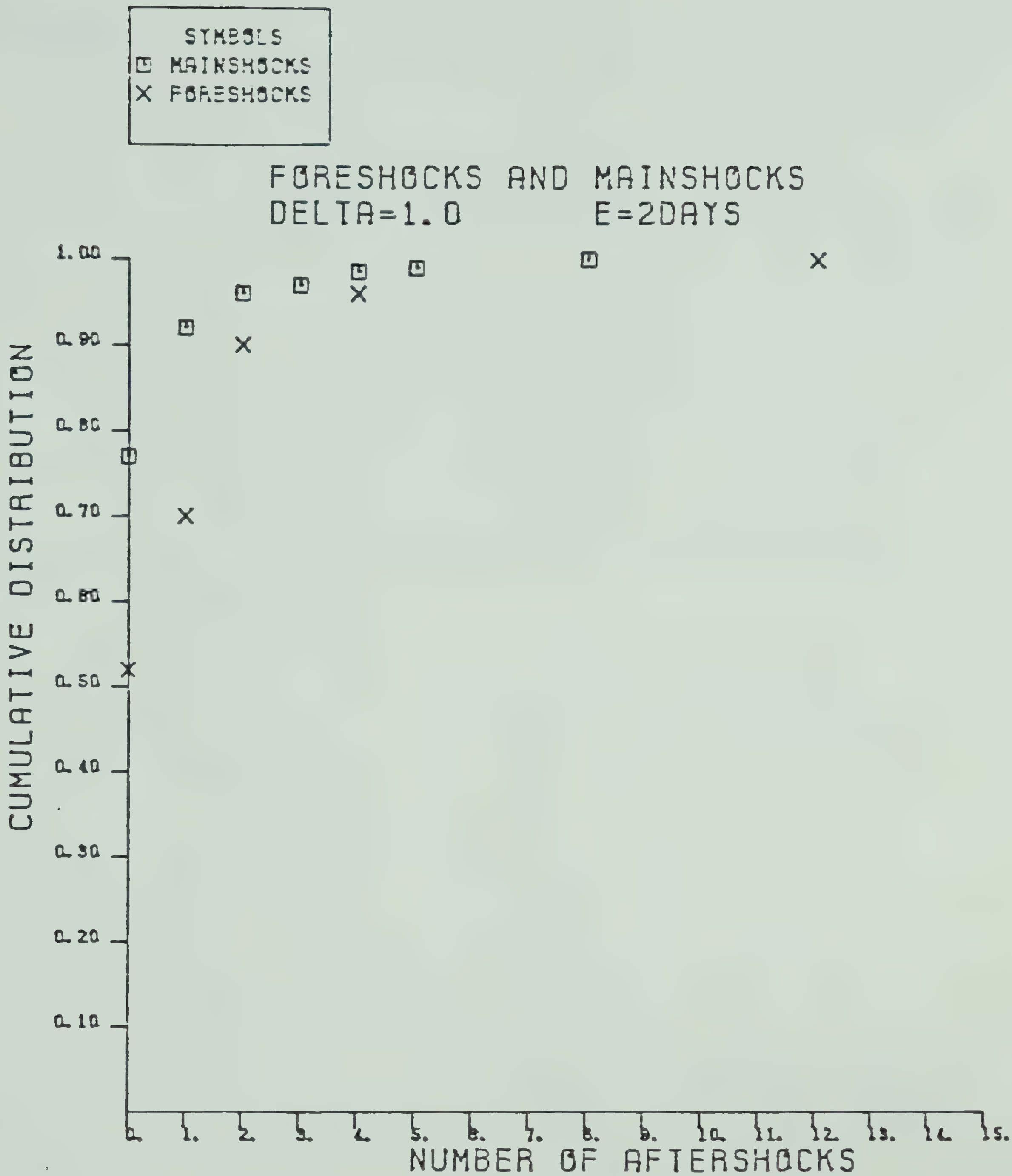


Figure 2.7 shows the normalized cumulative frequency distributions b(q) for foreshocks and mainshocks with Δ=1.0. Aftershocks were counted for 2 days.

2.7 Summary

The scaling law hypothesis seems to be supported, with the qualification that apparently the distribution of aftershocks is regionally dependent. Regional dependence, if present, increases the amount of data which is necessary for statistical studies of aftershock occurrence. The regional dependence may be related to changes in local tectonic style. Areas characterized by predominately extensional tectonics may show systematic differences in aftershock occurrence from areas characterized by predominately by compressionial tectonics.

It is suggested that, on the average, foreshocks generate larger numbers of aftershocks than do mainshocks. More data is needed before this observation can be tested.

The scaling law presented here as well as self-similarity in general have several implications for premonitory pattern "bursts of aftershocks". Anomalous bursts of aftershocks should be identifiable even for smaller magnitude mainshocks, provided the catalogue used is complete enough. The regional dependence of the scaling law may complicate the identification of what constitutes anomalous activity at a particular location. Keilis-Borok et al, (1980b) suggest that in the entire Southern California region anomalous bursts of aftershocks in the mainshock range of 5.5 to 6.4 are precursory to $M_0 \geq 6.5$ earthquakes, with a precursory time of the order of 3 years. If this is correct the scaling law and self-similarity suggest a

further question: Are anomalous bursts of aftershocks with smaller mainshock magnitudes precursory to intermediate magnitude events on a shorter time-distance scale? The Southern California catalogue is probably insufficiently complete to test this hypothesis at present.

3. Swarms and clusters of mainshocks in Southern California

"Figures often beguile me, particularly when I have the arranging of them myself; in which case the remark attributed to Disraeli would often apply with justice and force: 'There are three kinds of lies, lies, damn lies and statistics'."

Mark Twain
Autobiography,
Volume 1 page 246.

3.1 Introduction

Clustering of earthquakes in space and time is one of the most prominent features of seismicity. Phenomenology and statistical models of clustering are described in many studies (i.e. Utsu, 1972; Knopoff, 1971; Keilis-Borok et al., 1971; and Dziewonsky and Prozorov, 1981). Many of the proposed precursory seismicity patterns described in Chapter 1 involve recognition of abnormal clustering or anti-clustering of earthquakes. These include, for example: bursts of aftershocks (Keilis-Borok et al., 1980a,b); swarms (Caputo et al., 1977; McNalley, 1977; and Evison, 1977a,b); doughnut patterns (Kanamori, 1981); and seismic gaps of the second kind (for example, Mogi, 1979). The physical mechanism of clustering remains unclear: in particular, it is difficult to explain the interconnections between earthquakes over such large space and time intervals, which are suggested by some of the observed patterns. In the above

mentioned patterns no distinction was made between mainshocks, foreshocks and aftershocks, all were included.

Although there is no unambiguous physical definition for either a mainshock or an aftershock, the most obvious examples of earthquake clustering are mainshock - aftershock sequences. This work uses a very simple definition for mainshocks and aftershocks which will be discussed in the next section. The following questions arise: Do the sequences of obvious mainshock - aftershock clusters show a higher level of clustering? Are the "mainshocks", defined by the simple criteria, themselves clustered on a larger time - distance scale than is seen in the mainshock - aftershock clustering? If the mainshocks are clustered, are periods of abnormal clustering precursory to large earthquakes? Are precursory swarms and periods of quiescence observable in the sequence of the simple mainshock - aftershock clusters?

This chapter will discuss some of the statistical properties of the sequence of mainshocks in Southern California. The question of mainshock clustering is explored and, the clusters of mainshocks are described and juxtaposed with strong earthquakes. Part 1 uses a simple swarm model of clustering. In part 2 several clustering sequences are examined in more detail by means of cluster analysis.

Much of the research on this chapter was carried out at the Seismological Laboratory, California Institute of Technology while the author was assisting Prof. V. I. Keilis-Borok during his tenure there as Fairchild Scholar.

Part 1 of this chapter is based in part on the paper "Swarms of Main Shocks in Southern California" by V. I. Keilis-Borok, R. Lamoreaux, C. Johnson, and B. Minster.

3.2 On the definition of mainshocks and aftershocks

The question of the definition of an aftershock is a fundamental one. A conceptual definition of an aftershock is an event which is a direct result of the redistribution of energy resulting during failure of a previous shock. Stochastic models of earthquake sequences such as the branching model of Kagan and Knopoff (1981) mentioned in Chapter 1 present a conceptual problem in dividing earthquakes into foreshock - mainshock - aftershock sequences if a purely statistical definition is assumed in which mainshocks are defined simply as those events which are statistically independent of other events and aftershocks and foreshocks are defined as statistically dependent events. In such models all or most earthquakes will be interdependent. If one wishes to discuss mainshocks, aftershocks, and foreshocks some physical rather than statistical definition is needed. That is not the purpose of this work. The purpose of this chapter is to study clustering of earthquake sequences at intermediate time scales, time scales of the order of years. A simple definition of mainshocks and aftershocks is used to eliminate the obvious short term clustering in the

catalogue. The terms mainshock and aftershock used here should be considered in light of this.

In this work mainshocks and aftershocks were defined using the definition of aftershocks given in section 2.3.

$$ta-tm \le T(Mm);\ Ram \le R(Mm);\ Ma \le Mm \qquad 9$$

were t is the origin time and M is the magnitude; indices a and m indicate aftershock and mainshock respectively; Ram is the distance between the epicenters of a mainshock and its aftershock; T and R are empirical functions assumed after Gardner and Knopoff (1974) and Keilis-Borok et al. (1980b). This defintion is simple. What is used is a simple boxcar filter which is dependent on the magnitude of the largest event in a sequence.

Omori's law that the rate of aftershock generation falls off as the inverse of time results in a stong criticism of this approach. If Omori's law is correct for all time then the total expected number of aftershocks in time T for a given event will be

$$N = C\sum_{n=1}^{T} 1/n \qquad 10$$

The summation of 1/t does not converge. Essentially this implies that each event, if it generates aftershocks, generates an infinite tail of aftershocks. The question arises, how do we know that a present earthquake is not simply an aftershock of some earthquake which occurred in the perhaps distant past? Again at the root of this question

lies the exact question of what is the definition of an aftershock. After long periods of time the question of whether a earthquake is an aftershock of some previous event, which occurred at nearly the same location, or is perhaps due to changes in local conditions not related to the previous event becomes ambiguous. Perhaps the subsequent event could be more easily attributed to energy input into the system later in time than the previous events through some large scale driving force (i.e. plate tectonics). In such a case the event should not be termed an aftershock. Also, there is no guarantee that Omori's law is valid for very long time periods.

If Omori's law is assumed to be correct for all times an order of magnitude estimate of the number of aftershocks of, for example, all magnitude 6 or greater events can be easily made. For the last 45 years there have been on the average approximately one earthquake with a magnitude 6 or more per two years in the Southern California catalogue. Although there are great fluctuations from event to event, an estimate of C in Omori's law for a "typical" magnitude 6 earthquake can be made. Such an estimate gives an approximate value of C=2 days for aftershocks greater than magnitude 4. Assuming that we have a standard earthquake uniformly distributed every T_0 the expected number of "aftershocks" at time T which are generated by these events is given by

$$N = \sum_{n=1}^{L} \frac{C}{T + nT_o} \qquad 11$$

where T_o is the average recurrence period and L is some cutoff number of cycles indicative of the lifetime of a fault system. For T_o=2 years, C=2 days, L=1,000,000 and assuming T is small compared to nT_o the summation gives approximately 10 events per year. This number is significant when it is considered that there are on the average approximately 70 events/year with magnitudes greater than 4. If Omori's law is correct for all time of the order of 10% of these events are explanable as "aftershocks" of previous large events. These event will, however, appear to be statistically independent due to the different "mainshocks" to which they are associated and the long period of time from the "mainshock" to "aftershocks" of the same "mainshock".

An individual aftershock sequence might be modeled as following a time dependent Poisson distribution in which the mean number of events per unit time decays as 1/t. If such a sequence of events is observed after some period of time t_o (i.e. the first t_o days is not recorded) the series will appear random if time periods are observed which are small compared to t_o. For a Poisson process the variance of the distribution of events in any time interval Δt will be equal to the mean number of events in that time interval. If, for example, it is assumed that a change in the mean of 25% is significant, a significant change in the variance of the

distribution of events following the 1/t time dependence will not be noticable unless many time intervals longer than t/4 are taken, where t is the time from the initial shock. Also, long enough time intervals must be taken to adequately estimate the mean number of events if differences are to be noticable. Clearly we can not tell if a sequence of shocks occurring in a present catalogue are "aftershocks", using Omori's law, of some event occurring at a time t_0 before the start of the catalogue if t_0 is greater than the present length of the catalogue. Such sequences will appear to be random. We can not make any definite conclusion about properties of the sequences which show variations only at time scales of the order of or longer than the length of the available catalogues. The length of the Southern California catalogue is of the order of 50 years. If we assume that the seismicity of the last 50 years is representative of the preceding 50 years we can estimate the number of magnitude 4 "aftershocks" in the first year of the catalogue which are due to magnitude 6 events occurring in the 50 years which preceded the catalogue as approximately 4. This falls to less than 2 after 6 years. This small number of events will not significantly effect the statistics of the overall distribution of events.

The effects of the finite time windows in the aftershock definition can also be considered. Consider a magnitude 6+ earthquake which occurs at the beginning of the catalogue. The number of aftershocks missed, if Omori's law

is assumed to be true, due to the finite length of the aftershock time window can be estimated for a typical event. Assuming the same parameters as before and taking T(M) from table 2.1 gives an estimate that 3 to 5 aftershocks with magnitudes greater than 4 will be missed in the 50 year period of the catalogue. This small number will have little effect on the overall statistics of the entire sequence of earthquakes.

Consider another related question. It is well documented that large earthquakes on plate boundaries recur with similar magnitudes at nearly the same location. It has also been suggested that nearly all the strain release along segments of plate boundaries characterized by the occurrence of large earthquakes is released by the large earthquakes themselves. Approximate return times for the largest events in a region can be estimated independently of the seismicity patterns using constraints from plate tectonics. A review of this was given in section 1.4.3. Suppose that all earthquakes occurring on these segments are aftershocks of the series of largest earthquakes. The question arises, after what period of time will the probability of an earthquake being an aftershock of the last large event be equal to the probability of being an aftershock of all the preceding large events. Assuming the large earthquakes occur at regular intervals of T_0 and all generate aftershocks initially at the same rate the time is given by

$$T = T_0 / \Sigma \frac{1}{n}$$

where the sum is taken over all cycles. If 100000 cycles are taken T is approximately $T_0/14$. In Middle America the return times of large earthquakes have been estimated using plate tectonic considerations as 35 years (McNalley and Minster, 1981). This means that 2.5 years after the occurrence of a large event we are faced with the ambiguity of saying that an event which occurred in the epicentral region of the last large event is more likely to be an "aftershock" of one the earthquakes occurring in one of the distant past cycles of the strain release. The question of what is an aftershock of what becomes a unanswerable. Clearly, some other physical criteria is needed if the concept of an infinte tail of aftershocks is used.

In this chapter the possibility of long term clustering in earthquake sequences is studied. The time length of long term clustering which can be studied is restricted to periods which are short compared to the length of the catalogue. Nothing can be concluded about clustering which has time scales which are of the order of the length of the catalogue or longer. The Southern California catalogue spans approximately 50 years, this limits the study of "long term" clustering to periods which are a fraction of this. Here "long term" clustering will refer to clustering on time scales of 1 year to possibly 10 years.

A valid way to look for longer term clustering is to first filter out the short term variations which could possibly obscure the picture. This was the approach taken here. The purpose of the box car definitions for aftershocks was to remove the obvious short term clustering which occurs following a shock. In this study "mainshock" then refers to the largest event in one of these primary, box car defined clusters. "Aftershocks" and "foreshocks" are those events which occur respectively after and before the "mainshock" in the box car window. Foreshocks can have foreshocks and aftershocks can have aftershocks in this definition. The definitions here are obviously not unique. Mainshocks defined by the above criteria for a particular choice of R(M) and T(M) will not necessarily be mainshocks by another criteria.

The boxcar filters used here are strong. Even if the 1/t dependence suggested by Omori's law holds for very long periods of time most "aftershocks" occurring in the length of the catalogue will fall within the boxcar windows. The few aftershocks which are missed will be widely spread in time. These "missed aftershocks" will not affect the overall statistics of the distribution of residual events.

3.3 Statistics of mainshocks

The Southern California catalogue 1932 through 1980 (Hileman et al., 1973; Freidman et al., 1976; and Subsequent Preliminary Epicenter Listings) provided the data for the study. Two separate mainshock catalogues were created using the two versions of the thresholds T(M) and R(M) given in Tables 2.1 and 3.1. The thresholds in Table 2.1 were suggested in Gardner and Knopoff, (1974); Table 3.1 gives a simplified version, used in the global test of premonitory clustering of earthquakes (Keilis-Borok et al., 1980b).

The number of mainshocks in different time and magnitude ranges is shown in Tables 3.2 and 3.3. A map of mainshocks with $Mm \geq 5.0$ (using the threshold of Table 2.1) is given in Figure 3.1. It is interesting to notice the general quiescence in the occurrence of mainshocks from 1968 to 1978 (Figure 3.2).

3.4 Clustering of mainshocks

Knopoff and Gardner (1974) found that the distribution of mainshocks, using the definition of table 2.1, in short time intervals for the whole of Southern California is not distinguishable from a Poisson distribution. However, this does not eliminate the possibility that the sequences of mainshocks in smaller areas deviate from the Poisson distribution and, in particular, that clusters of mainshocks occur. In order to test this, the distribution of the number

Simplified windows for the identification of aftershocks (after Keilis-Borok et al., 1980b).

M	R(km)	T(days)
3.0 - 3.9	50.0	46.0
4.0 - 4.4	50.0	91.0
4.5 - 5.4	50.0	182.0
5.5 - 6.4	50.0	365.0
> 6.4	50.0	730.0

Table 3.1 gives a simplified version of R(M) and T(M) used in the identification of aftershocks (after Keilis-Borok et al., 1980b).

Table 3.2 is a tabulation of the number of mainshocks in different time and magnitude intervals. Mainshocks are defined by table 2.1. Each column indicates the number of events from the magnitude interval heading the column to but not including the magnitude heading the next column. The time interval is indicated with the last year not inclusive. The average magnitude is calculated for magnitudes greater than 3.5. The stability of the average magnitude is an indication of catalogue completeness for magnitudes above 3.5.

Year	Av. Mag.	3.0	3.4	3.6	3.8	4.0	4.2	4.4	4.6	4.8	5.0	5.2	5.4	5.6	5.8	6.0	6.2	>6.4	Total
1932-1935	3.97	131	80	6	0	32	3	21	1	0	5	0	1	0	0	2	1	4	287
1935-1938	3.90	165	77	0	0	51	1	22	1	0	8	0	0	0	0	1	0	0	326
1938-1941	4.01	132	91	0	1	56	0	31	3	1	11	0	5	1	0	2	0	2	336
1941-1944	3.94	118	86	3	1	54	4	15	0	2	5	1	5	0	2	2	0	0	298
1944-1947	4.01	103	47	39	23	25	16	8	7	4	4	3	1	2	0	0	1	0	283
1947-1950	4.06	138	42	41	21	17	12	9	12	8	2	3	1	2	1	0	1	1	311
1950-1953	4.00	159	54	29	25	18	8	7	3	4	3	2	2	1	1	1	0	1	318
1953-1956	4.12	149	35	27	25	24	17	9	3	4	2	5	2	2	1	2	2	0	309
1956-1959	4.00	148	46	31	22	17	14	6	12	4	1	2	0	0	1	1	0	1	306
1959-1962	4.04	149	48	32	29	31	20	16	8	5	2	6	1	1	0	0	1	0	359
1962-1965	4.10	151	60	28	26	35	16	21	9	10	12	1	1	1	2	0	0	0	373
1965-1968	4.00	132	56	32	17	25	13	14	5	7	2	3	1	1	0	0	1	0	309
1968-1971	3.96	212	63	48	23	21	5	9	2	5	3	4	2	2	1	0	0	1	401
1971-1974	3.87	235	70	44	26	6	9	4	3	3	3	0	0	0	1	0	0	1	426
1974-1977	3.97	250	67	40	23	18	12	8	3	4	3	1	0	0	0	0	0	0	465
1977-1980	3.97	230	63	40	23	15	9	6	4	4	2	2	1	0	2	0	0	1	402
1980-1981	4.00	72	25	18	11	6	1	6	2	0	1	0	1	1	0	0	0	2	146
Total		2674	1020	454	313	474	163	221	84	64	70	35	25	14	12	11	7	14	5655

Table 3.3 gives the same tabulation as table 3.2 but uses the mainshock definition of table 3.1

Year	Av. Mag	3.0	3.4	3.6	3.8	4.0	4.2	4.4	4.6	4.8	5.0	5.2	5.4	5.6	5.8	6.0	6.2	≥ 6.4	Total
1932-1935	4.01	90	67	5	0	29	3	21	1	0	4	0	1	0	0	2	1	4	228
1935-1938	3.93	123	59	0	0	48	1	19	1	0	7	0	0	0	0	1	0	0	259
1938-1941	4.03	86	78	0	1	52	0	26	3	1	11	0	5	1	0	2	0	2	268
1941-1944	4.00	89	66	1	1	41	4	13	0	2	6	1	4	0	2	2	0	1	233
1944-1947	4.04	73	36	35	17	21	14	8	6	4	4	3	1	2	0	0	1	0	225
1947-1950	4.08	102	34	34	20	15	10	7	12	5	2	3	1	2	1	0	1	1	250
1950-1953	4.03	114	52	26	23	14	10	7	3	3	4	2	2	1	1	2	0	1	265
1953-1956	4.15	107	35	22	20	20	14	8	3	4	2	5	2	2	1	2	2	0	249
1956-1959	4.03	91	41	26	21	16	13	6	12	4	1	2	0	0	1	1	0	1	236
1959-1962	4.08	112	47	25	23	24	17	14	8	5	2	6	1	1	0	0	1	0	286
1962-1965	4.12	109	43	23	22	30	13	17	9	9	11	1	1	1	2	0	0	0	291
1965-1968	4.02	96	43	26	14	21	12	14	5	6	1	3	1	1	0	0	1	0	244
1968-1971	3.97	160	50	40	19	21	5	8	0	4	3	3	2	2	1	0	0	1	319
1971-1974	3.89	180	54	37	22	21	6	8	3	3	2	0	0	0	1	0	0	1	338
1974-1977	4.01	168	53	28	28	22	17	12	8	4	4	2	1	0	0	0	0	0	347
1977-1980	3.99	173	58	34	22	14	8	7	3	3	2	2	1	0	2	0	0	1	330
1980-1981	4.07	60	18	16	9	5	1	6	2	0	1	0	1	1	0	1	0	2	123
Total		1933	834	378	262	414	148	201	79	57	67	33	24	14	12	13	7	15	4491

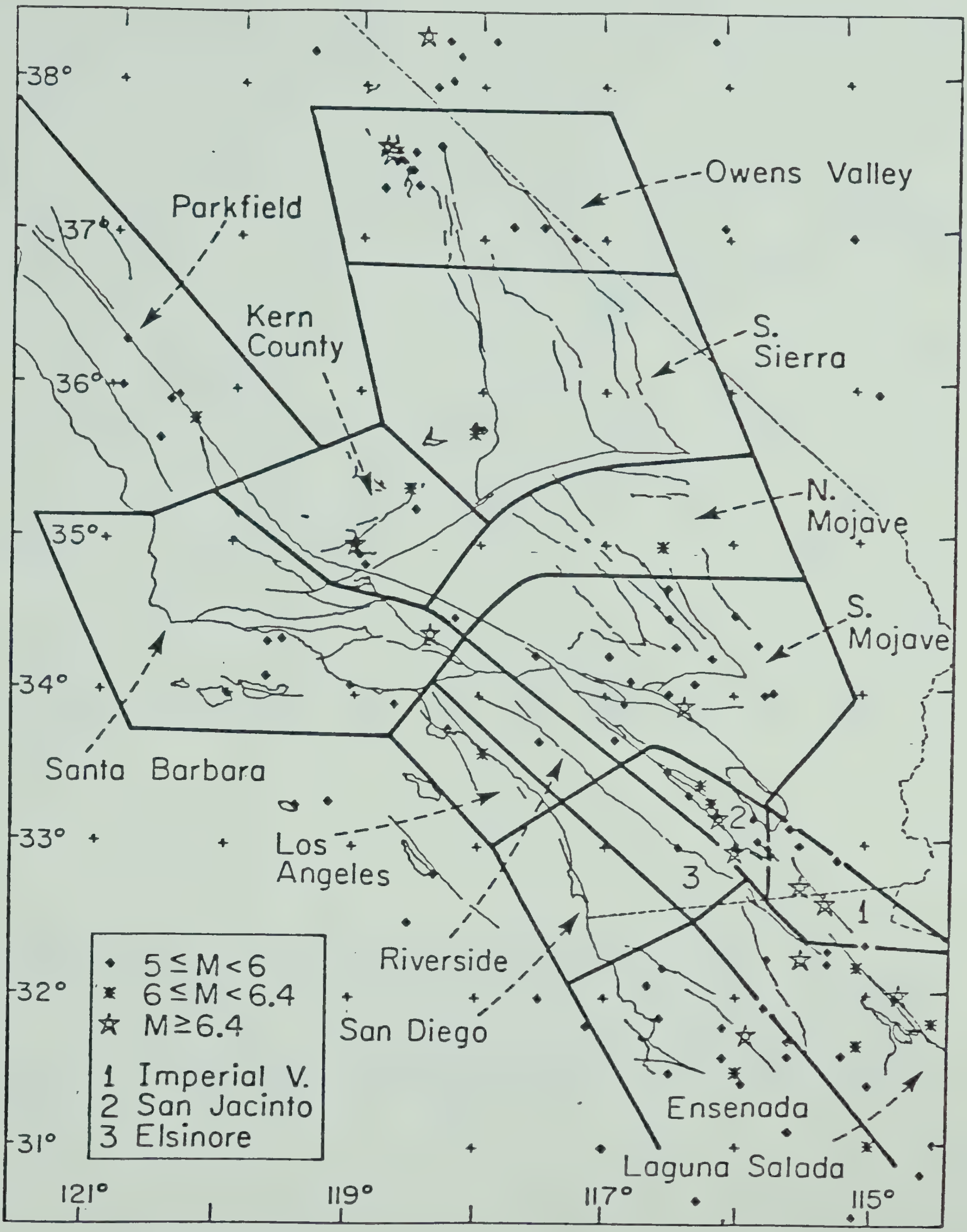


Figure 3.1 shows a map of mainshocks for Southern California with magnitudes greater than 5.0. The aftershock thresholds of table 2.1 were used. The boundaries of the 14 subregions are also given.

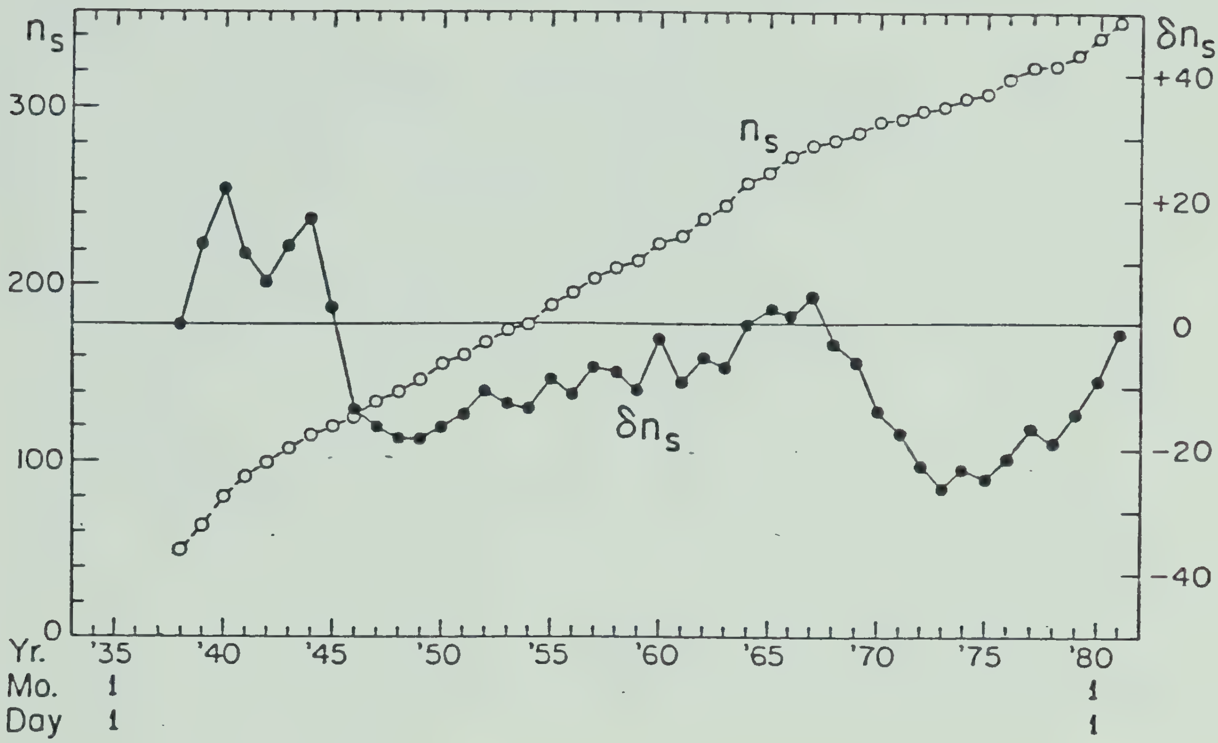


Figure 3.2 gives a plot of the cumulative number of mainshocks Ns, M≥4.0, from the beginning of the catalogue. δNs gives the difference between the cumulative number of mainshocks and the number predicted from the average number of events per year over the time interval ending 6 years before the date given. Note the general quiescence in the occurrence of mainshocks from 1968 to 1978.

of mainshocks in rectangular boxes of size Δx, Δy, s were considered. The parameters Δx, Δy, and s refer to the dimensions of the box in latitude, longitude and time respectively.

Let n be the number of earthquakes in one of these boxes and let N be the total number of mainshocks in a single rectangle Δx by Δy for the entire period of time spanned by the catalogue. The two-dimensional histograms P(n,N) were tabulated. P(n,N) gives the number of times all rectangles Δx by Δy, which contained N events over the entire catalogue, had n events in the time intervals of duration s. The following values of parameters were assumed: $\Delta x = \Delta y = .4°$, s= 1, 3 and 6 years; minimal magnitude M_0 = 3.5 and 4.0. Two examples of these histograms are given in Tables 3.4 and 3.5.

The distributions P(n,N) were compared to the expected Poisonian distributions, with the same average number of events per time window, using the Chi-squared criteria (e.g., Cramer, 1946). The values of P in each line were grouped in such a way that each group had at least 5 members. The results are given in Table 3.6. Table 3.6 shows that the deviations between the observed distributions and the corresponding Poisson distributions are significant. In all but two cases the hypothesis that the distributions P(n,N) follow the Poisson law can be rejected at less than the .001 level of significance. Additional confirmation of the existence of clustering is that for n = 0 the observed

values of P are systematically greater than frequencies predicted by Poisson distributions.

3.5 Clustering of mainshocks without foreshocks

In the preceding section foreshocks were included in the analysis. The analysis demonstrated that the sequences of mainshocks plus foreshocks as defined by Tables 2.1 and 3.1 show significant deviations from Poisson distributions. The deviations were such as to suggest a significant degree of clustering, however the significance of the contribution of the foreshocks was not considered. The following question remains: is the observed clustering due simply to the presence of foreshocks?

Gardner and Knopoff (1974) show that if aftershocks and foreshocks are eliminated from the catalogue by means of the criteria of Table 2.1, the distribution of numbers of mainshocks in 10 day intervals over all of Southern California is indistinguishable from a Poisson distribution.

To test for local clustering of mainshocks with foreshocks removed, the Chi-square analysis of the preceding section was repeated on the catalogue of mainshocks as defined by Table 2.1 excluding foreshocks. P(n,N) was constructed for various values of Δx, Δy, s, and M_0. The results are given in Table 3.7.

The sequence of earthquakes (including aftershocks and foreshocks) shows a high degree of clustering. The removal

Number of Events in a Cell for the Entire Catalog, N	Number of Events in a Cell per 6 Years, n												
	0	1	2	3	4	5	6	7	8	9	10	11	>11
0	768	0	0	0	0	0	0	0	0	0	0	0	0
1	301	43	0	0	0	0	0	0	0	0	0	0	0
2	203	56	5	0	0	0	0	0	0	0	0	0	0
3	82	31	7	0	0	0	0	0	0	0	0	0	0
4	86	34	14	2	0	0	0	0	0	0	0	0	0
5	57	53	8	2	0	0	0	0	0	0	0	0	0
6	62	40	22	4	0	0	0	0	0	0	0	0	0
7	10	11	1	0	2	0	0	0	0	0	0	0	0
8	23	17	12	3	0	1	0	0	0	0	0	0	0
9	19	14	11	2	1	0	0	1	0	0	0	0	0
10	13	19	12	2	0	1	1	0	0	0	0	0	0
11	12	7	5	5	3	0	0	0	0	0	0	0	0
12	10	11	4	4	0	2	0	1	0	0	0	0	0
13	11	12	6	7	0	4	0	0	0	0	0	0	0
14	12	10	7	4	4	1	2	0	0	0	0	0	0
15	11	13	9	8	4	1	1	0	1	0	0	0	0
16–18	6	22	16	8	8	2	0	2	0	0	0	0	0
19–21	9	12	10	9	6	5	4	0	1	0	0	0	0
22–24	4	4	3	1	1	0	0	0	1	0	1	0	1
25–27	3	1	4	2	2	1	1	0	2	0	0	0	0
28–30	1	6	4	9	4	1	2	3	1	0	0	0	1
31–33	5	4	5	8	2	5	1	4	2	3	1	0	0
34–36	3	0	4	9	6	1	2	1	1	2	3	0	0
37–39	2	1	2	6	1	1	4	3	2	0	1	1	0
40–42	0	0	1	7	0	1	3	0	1	1	2	0	0
>42	0	1	3	3	5	2	8	2	3	4	2	4	3

Table 3.4 gives an example of P(n,N) using the mainshock definition of table 2.1 and mainshocks with M≥3.5.

Number of Events in a Cell for the Entire Catalog, N	Number of Events in a Cell per 6 Years, n												
	0	1	2	3	4	5	6	7	8	9	10	11	>11
0	784	0	0	0	0	0	0	0	0	0	0	0	0
1	322	46	0	0	0	0	0	0	0	0	0	0	0
2	189	56	3	0	0	0	0	0	0	0	0	0	0
3	110	40	10	0	0	0	0	0	0	0	0	0	0
4	83	40	11	2	0	0	0	0	0	0	0	0	0
5	68	62	14	0	0	0	0	0	0	0	0	0	0
6	35	23	11	3	0	0	0	0	0	0	0	0	0
7	33	25	10	1	2	1	0	0	0	0	0	0	0
8	13	18	6	2	1	0	0	0	0	0	0	0	0
9	16	16	12	2	2	0	0	0	0	0	0	0	0
10	13	2	3	3	2	1	0	0	0	0	0	0	0
11	13	12	8	3	3	0	1	0	0	0	0	0	0
12	13	11	8	3	1	4	0	0	0	0	0	0	0
13	6	11	8	5	0	2	0	0	0	0	0	0	0
14	11	16	16	5	7	0	1	0	0	0	0	0	0
15	7	16	7	3	4	1	0	1	1	0	0	0	0
16–18	8	9	8	9	3	2	0	1	0	0	0	0	0
19–21	11	15	9	3	8	3	3	4	1	0	0	0	0
22–24	3	2	3	3	2	0	1	3	0	0	0	0	0
25–27	4	8	9	9	7	3	2	3	2	0	1	1	0
28–30	4	3	5	6	4	3	3	0	2	2	0	0	0
31–33	3	2	9	9	4	3	2	1	2	4	1	0	0
34–40	0	0	0	0	0	0	0	0	0	0	0	0	0
>42	0	2	6	8	3	5	7	6	3	4	2	1	1

Table 3.5 gives an example of $P(n,N)$ using the mainshock definition of table 3.1 and mainshocks with $M \geq 3.5$.

M	R(M), T(M) after table	s	x	n	P
$m \geq 4.0$	2	6 years	20.22	15	.84
$m \geq 3.5$	2	6 years	66.27	32	.999
$m \geq 4.0$	1	6 years	27.27	18	.93
$m \geq 3.5$	1	6 years	74.66	34	.999
$m \geq 4.0$	2	3 years	43.69	12	.999
$m \geq 3.5$	2	3 years	95.08	35	.999
$m \geq 4.0$	2	1 year	14.1	7	.999
$m \geq 3.5$	2	1 year	27.6	17	.999

Table 3.6 gives a Chi-squared test of the two dimensional histograms. In the table s gives the time window used in counting the number of events in a .4 by .4 degree square of latitude and longitude, n gives the number of degrees for the Chi-squared test, x gives the value of χ^2, and P is the probability that a random variable which follows the Chi-square distribution with n n degrees of freedom is less than or equal to x.

Table 3.7 gives a Chi-squared comparison of the two dimensional histograms P(n,N), excluding foreshocks, to Poisson distributions. In the table s gives the time window used in counting the number of events, M the minimal magnitude, n the number of degrees of freedom, z the value of x^2, p the probability that a random variable which follows the Chi-square distribution with n degrees of freedom is less than or equal to z. "I" gives the number of ranges which had variances greater than the variances predicted by a Poisson process and q gives the probability of i or more of I independent measurements being too large assuming the probability of measuring a variance which is too large is the same as the probability of measuring a variance that is too small. Clustering is indicated by large values of p and small values of q.

Chi-squared test of distribution of mainshocks without foreshocks to a Poissonian distribution.

Area divided into .4 degree by .4 degree squares

s	M	n	z	p	I	i	q
1 year	3.5	14	10.18	.250	30	16	.572
1 year	4.0	6	7.95	.758	18	5	.996
3 years	3.5	38	70.20	.998	29	23	.001
3 years	4.0	12	15.62	.791	17	12	.072
6 years	3.5	32	48.04	.966	22	15	.067
6 years	4.0	14	18.33	.808	16	12	.038

Area divided into .7 degree by .7 degree squares

s	M	n	z	p	I	i	q
100 days	3.5	8	6.64	.424	23	10	.798
100 days	4.0	1	.11	.260	15	5	.941
1 year	3.5	23	29.92	.848	23	15	.105
1 year	4.0	8	12.88	.884	15	7	.696
2 years	3.5	31	37.05	.790	22	17	.008
2 years	4.0	15	21.10	.866	15	7	.696
3 years	4.0	14	24.28	.958	15	11	.059
3 years	4.5	5	4.48	.517	7	3	.773
4 years	3.5	17	26.68	.937	16	10	.227
4 years	4.0	12	17.40	.865	11	8	.113
4 years	4.5	5	14.44	.987	4	3	.312
5 years	4.0	7	10.36	.830	7	5	.227
5 years	4.5	5	9.52	.870	5	4	.187
6 years	4.0	9	17.87	.963	8	6	.144
6 years	4.5	6	8.86	.818	5	4	.187

Area divided into 1 degree by 1 degree squares

s	M	n	z	p	I	i	q
42 days	3.5	6	3.47	.252	25	12	.654
100 days	3.5	13	25.77	.982	27	19	.026
100 days	4.0	5	3.16	.325	17	4	.994
1 year	3.5	30	55.53	.997	24	20	.001
1 year	4.0	14	16.15	.696	17	7	.834
2 years	3.5	29	44.00	.989	24	22	.0001
2 years	4.0	17	26.38	.932	14	11	.029
2 years	4.5	5	2.74	.260	7	4	.500
3 years	3.5	25	49.11	.997	23	20	.0002
3 years	4.0	11	19.24	.943	12	9	.073
3 years	4.5	6	9.46	.851	6	4	.344
4 years	3.5	10	22.94	.989	22	17	.008
4 years	4.0	8	13.79	.913	11	8	.113
4 years	4.5	5	6.57	.745	4	3	.312
5 years	4.0	10	17.39	.934	12	12	.0002
5 years	4.5	5	8.66	.877	6	6	.016
6 years	4.0	7	11.60	.886	10	8	.055
6 years	4.5	4	8.86	.935	6	5	.109

Whole Southern California region

s	M	n	z	p
10 days	3.5	4	5.07	.720
15 days	4.0	4	.73	.052
100 days	3.5	11	8.15	.300
100 days	4.0	7	6.19	.482
1 year	4.5	4	2.66	.384
1 year	5.0	3	2.13	.454

of aftershocks and foreshocks in the manner described here and by Gardner and Knopoff (1974) can be viewed as a strong filtering process which removes the obvious mainshock - aftershock and mainshock - foreshock clusters from the sequence. This shifts the distribution of remaining events (mainshocks) away from a clustered distribution and towards a uniform distribution.

A distribution which shows clustering will have a larger variance than a Poisson distribution, while a distribution which tends towards a uniform distribution will have a smaller variance. In terms of P(n,N) this means that clustered distributions should generally have systematically more occurrences of cases in the tails of the distribution. In particular P(0,N) has an increased probability of being greater than that predicted by a Poisson distribution. For a uniform distribution the effect is the opposite.

As a further test for clustering the following was considered: for the various values of Δx, Δy, s and M_0 the resulting histograms P(n,N) were grouped into ranges of N such that the ranges contained more than 25 cases (this was relaxed to 16 cases for $s \geq 4$ years due to the decrease in the number of cases). For each range N the variances of the distributions P(n,N) were compared to those predicted by a Poisson distribution. In Table 3.7 "I" gives the number of ranges for a given Δx, Δy, s and M_0 and "i" gives the number of ranges which had variances greater than the variance predicted for a Poisson distribution. In Table 3.7 q gives

the probability of i or more of I independent measurements being too large assuming the probability of measuring a variance which is too large is the same as the probability of measuring a variance which is too small. A comparison of P(0,N) to the corresponding values predicted for a Poisson distribution gives almost identical results. Clustering is indicated by large values of p and small values of q.

For the whole Southern California region with s=10 days, 15 days, and 100 days and 1 year in the chosen magnitude ranges the results are in agreement with Gardner and Knopoff (1974). The distributions appear to follow a Poisson distribution.

For Δx and $\Delta y = .4^\circ$ The degree of clustering is significantly less than the case which includes foreshocks. The hypothesis that the distributions follow a Poisson distribution can not be rejected at the 95% level (except for $M_0 \geq 3.5$, s=3 and 6 years). There is, however, a suggestion of weak clustering for values of s of 3 and 6 years for all cases. Apparently, a significant amount of the observed clustering in the previous section was due to foreshocks. The tabulated cases suggest that differences between the distributions and Poisson distributions becomes progressively more significant as Δx, Δy, and s increase. This is indicated by increasing values of p and decreasing values of q as Δx, Δy, and s increase.

These results support the hypothesis that earthquake sequences are clustered on longer time and distance scales

than is indicated by mainshock - aftershock clustering, although the clustering is weaker than that suggested by the previous section. This clustering is evident on time scales and distance scales which are an order of magnitude larger than the time and distance windows used to eliminate aftershocks and foreshocks.

The failure to detect clustering on distance scales of the order of the entire Southern California region is not surprising. The catalogue of mainshocks will be dominated in number by events with magnitudes near the minimum magnitude cut-off. The statistical properties of the ensemble will be determined by this lowest magnitude range, larger events will not contribute significantly. Clustering is suggested to be a local effect. The time and distance scales of the clustering depends on the magnitude range considered. For the mimimum magnitudes considered here this range is an order of magnitude less than the size of the entire Southern California region. By using the entire Southern California region a superposition of several clustered distributions is obtained. The superposition of many clustered distributions may appear random. The clustering of mainshocks is only evident in time windows of the order of years. Due to the short duration of the catalogue (49 years) it is not possible to test for long-period, non-local clustering in the entire catalogue.

The results presented here suggest that as the minimial magnitude is increased clustering becomes evident at larger

distance - time scales. Unfortunately the duration of the catalogue is insufficient to test this hypothesis. For M≥5.0 time windows of 10 to 15 years might be needed. This would allow only 3 or 4 time intervals using the available catalogue.

The hypothesis that there are related events in the catalogue over long periods of time and large distances, such as is suggested by the stochastic branching models of earthquake occurrence, is not in conflict with the observations presented here. If, for example, the Kagan and Knopoff branching model, in which all shocks can be thought of as "children" of those shocks occurring before them in time, is correct no length of box car would be sufficient to eliminate all dependent events. The Kagan and Knopoff model was discussed in Section 1.11. In such a model clustering would always be observed at time scales longer than the length of the boxcar. The short length of available catalogues procludes the possibility of testing for the presence of such long term clustering.

3.6 Part 1: Quiescence, swarms of mainshocks and strong earthquakes

3.6.1 Hypothesis

The relevance of the clustering of earthquakes to the process of preparation of a strong earthquake has been reported in many studies. Some long-term premonitory seismicity patterns are directly formed by different types of clusters. The review of these patterns can be found in Chapter 1 as well as in Keilis-Borok et al. (1980b) and Kanamori (1981). Some long-term patterns (i.e. bursts of seismicity) have been shown to give good results in retrospective prediction tests. However, they do not indicate the place of the coming earthquake within the region. They indicate only a time-window (3-4 years for the bursts of seismicity) as do most other long-term patterns.

Some other clustering-related premonitory patterns, which have been suggested, imply the following sequence: quiescence → activation and/or abnormally large clusters → strong earthquake. The several suggested patterns of this type use different definitions of quiescence, activation and clusters. A review of these patterns is given in Chapter 1. These patterns are designed to indicate not only the time, but also the place of the coming large earthquake. However, so far they do not show "acceptable scores" even in retrospective prediction due to an excessive number of false alarms.

Many abnormally large clusters, encountered in the study of the these premonitory patterns, are formed by a mainshock and its aftershocks (the aftershocks of only the

strongest earthquakes are eliminated from consideration in the definition of some patterns). In the following sections the hypothesis that the mainshocks themselves form a similar sequence (quiescence $\rightarrow$ activation and/or clustering) before a strong earthquake will be tested.

3.6.2 Data analysis -- general scheme

The above mentioned hypothesis was considered in the following way.

(i) Southern California was divided in 14 areas, shown in Figure 3.1. These areas represent the zones of major faults; the location of these faults was taken from Hileman et al. (1973). Obviously the division, shown on Figure 3.1 is by no means unique, especially where the faults are dense and not separated by zones of relatively low seismicity.

(ii) The lists of mainshocks in each area were checked to see if the sequence {quiescence $\rightarrow$ abnormal clustering $\rightarrow$ strong earthquake} occurred. For this purpose the following functions were tabulated:

$N(t,M_0)$ -- the number of mainshocks in a sliding time window from $(t-s)$ to t, with magnitude $Mm \geq M_0$;

$R(t,M)$ -- maximal number of mainshocks (in the same time-window and magnitude range) with epicenters that can be placed within a rectangular cell Δx by Δy;

$\bar{N}(t,M_0)$ -- average value of $N(t,M_0)$ for the time interval from the beginning of the catalogue to t.

Swarms of mainshocks are defined by the conditions

$$R(t) \geq \max\{1/2N(t), C\};\ N(t) \geq \bar{N}(t) \qquad 13$$

This same definition was used in Caputo, et al. (1977) on earthquakes in Italy, however the Italian study did not separate mainshocks and aftershocks.

(iii) Functions $N(t)$, $N(t)/\bar{N}(t)$ and swarms were juxtaposed with the time of occurrence of strong ($Mm \geq 6.4$) mainshocks in each area. The hypothesis implies that strong earthquakes are preceded by a swarm (which is one of the types of abnormal clusters), and/or by a maximum of $N(t)$ or $N(t)/\bar{N}(t)$, which are the measures of activation; either one has to be preceded by a mimimum of $N(t)$ or $N(t)/\bar{N}(t)$, which would correspond to quiescence. Quiescence and activation are not formally defined here.

3.6.3 Data analysis -- results

The catalogue of mainshocks, using the thresholds $R(M)$ and $T(M)$ of Table 3.1, was analyzed. To secure the completeness of the catalogue the minimum magnitude threshold M_0 was chosen as 4.0 except for regions of low activity. This more or less predetermined the following choice of parameters: $\Delta x = \Delta y = .4°$, $s = 6$ years, $C=4$. If at the end of the catalogue the average number of mainshocks $\bar{N}$ in the 6-year windows was less than 6, the threshold M_0 was lowered sufficiently to obtain an $\bar{N}$ of from 6 to 8. These parameters are not claimed to be optimal. The results of the analysis are represented on figures 3.3 to 3.10. The strong

(Mm ≥ 6.4) mainshocks are listed for convenience in Table 3.8.

Vertical lines indicate the moment of strong (M ≥ 6.4) mainshocks. The numbers on the upper time scale indicates the year of occurrence and magnitude of the strong mainshocks outside the area represented on the plot. This makes it easier to check whether the occurrence of the pattern in one area is connected to the strong mainshocks in another area.

Values of $N(t)$, $N(t)/\bar{N}(t)$ and $R(T)$, according to the definition of these functions, refer to the end of the sliding 6-year time window. This window was moved in one year steps. The starting time for each area was shifted in such a way that the last strong mainshock falls at the beginning of a time window (the situation just before the strong mainshock is given in the window preceding the strong event). Accordingly, to avoid misunderstanding, the vertical line indicating the moment of the strong event was displaced by ~1 mm to the right. The actual moment is indistinguishable from the preceding time mark. The values of R are indicated only for swarms, each double circle signifies the occurrence of a swarm. It should be noted, however, that consecutive double circles often correspond to one and the same swarm which is concentrated on a shorter time interval than 6-years. With the choice of 6-year intervals we could estimate $N(t)$, $\bar{N}(t)$ and $R(t)$ only after 1937 as the catalogue begins in 1932. The estimates of $\bar{N}(t)$

Year	Month	Day	Latitude	Longitude	Magnitude	Region
1934	1	30	38.33	-118.50	6.6	Owens Valley
1934	12	30	32.25	-115.50	6.5	Laguna Salada
1934	12	31	32.00	-114.75	7.1	Laguna Salada
1940	5	19	32.73	-115.50	6.7	Imperial Valley
1942	10	21	32.97	-116.00	6.5	Elsinore
1948	12	4	33.93	-116.38	6.5	Southern Mojave
1952	7	21	35.00	-119.02	7.2	Kern County
1956	2	9	31.75	-115.92	6.8	Ensenada
1968	4	9	33.19	-116.13	6.4	San Jacinto
1971	2	9	34.41	-118.40	6.4	Santa Barbara
1979	10	15	32.61	-115.32	6.6	Imperial Valley
1980	5	25	37.56	-118.79	6.5	Owens Valley

Table 3.8 gives a listing of large events, $M \geq 6.4$, in Southern California.

become stable much later. Negative focal depth is indicated in the catalogue for some earthquakes (most of them have epicenters in the Owens Valley and Imperial Valley areas). They were elliminated from the calculations represented in Figures 3.3 to 3.10. The results obtained by including these earthquakes do not differ significantly from those presented.

3.6.4 Summary of the results

The areas considered can be divided into two groups. The first group consists of 7 areas for which a lower threshold magnitude of $M_0 = 4$ could be maintained (Figures 3.3 to 3.6). The data for this group seem to confirm the expected pattern. The three last strong earthquakes -- Owens Valley, 1980; Imperial Valley, 1979; and San Fernando, 1971 -- are preceded first by quiescence (represented by clear minimums of $N(t)$ and $N(t)/\bar{N}(t)$ and then by the swarms. The San Miguel earthquake, 1956, is preceded by a swarm, but not by a pronounced minimum in $N(t)/\bar{N}(t)$. $N(t)$ does show a minimum from 1941 to 1952, however $\bar{N}(t)$ had not achieved a stable value. For the 1948 Desert Hot Springs earthquake the results are negative. It was preceded by activation (an increase in $N(t)$) but not by a swarm or quiescence. Swarms were not diagnosed in the S. Mojave area at any time. For the Parkfield area the results are positive in the sense that neither the quiescence → swarm pattern or strong earthquakes occurred there (there were no false alarms). The

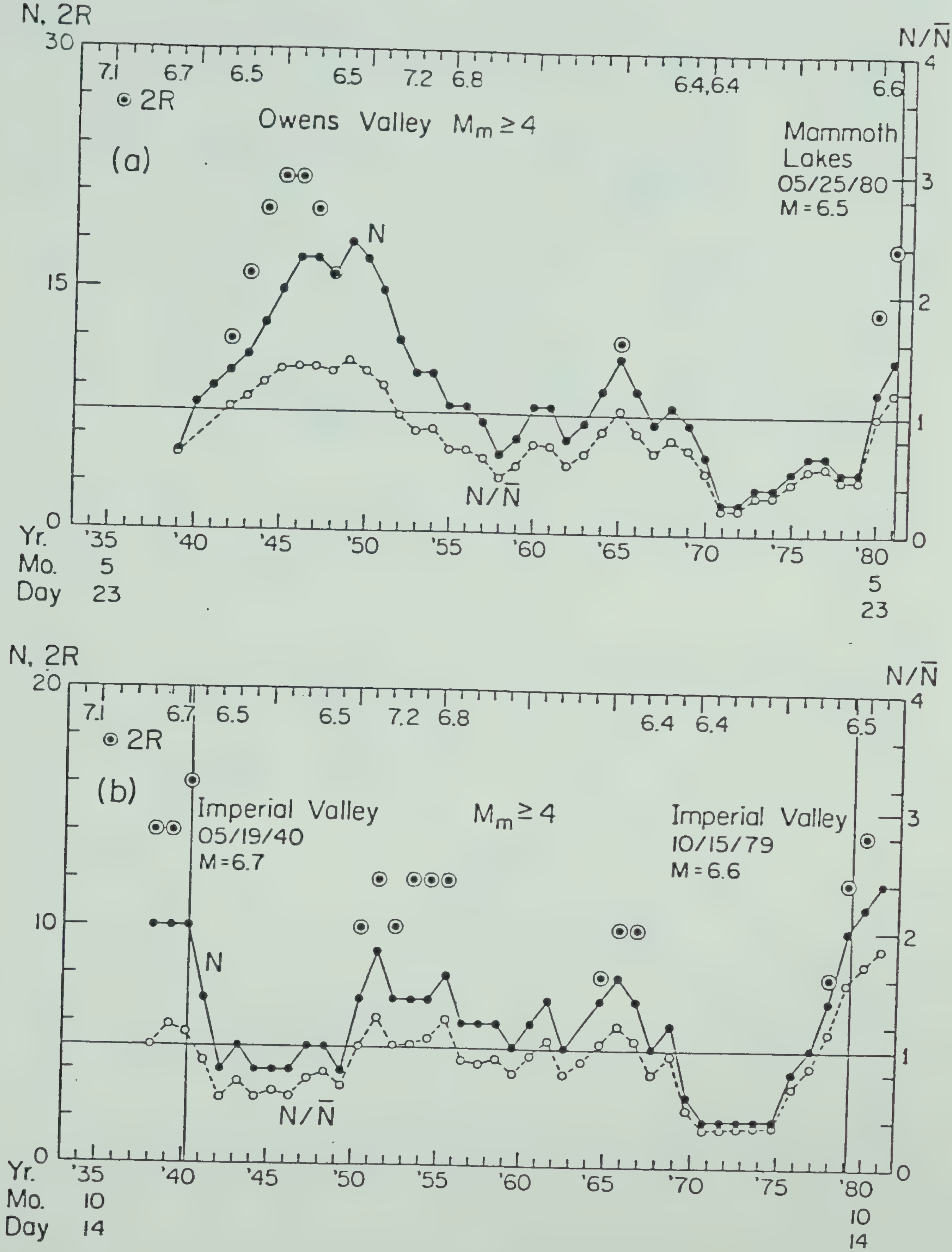


Figure 3.3 gives the test of the quiescence → activation hypothesis for the Owens Valley and Imperial Valley regions.

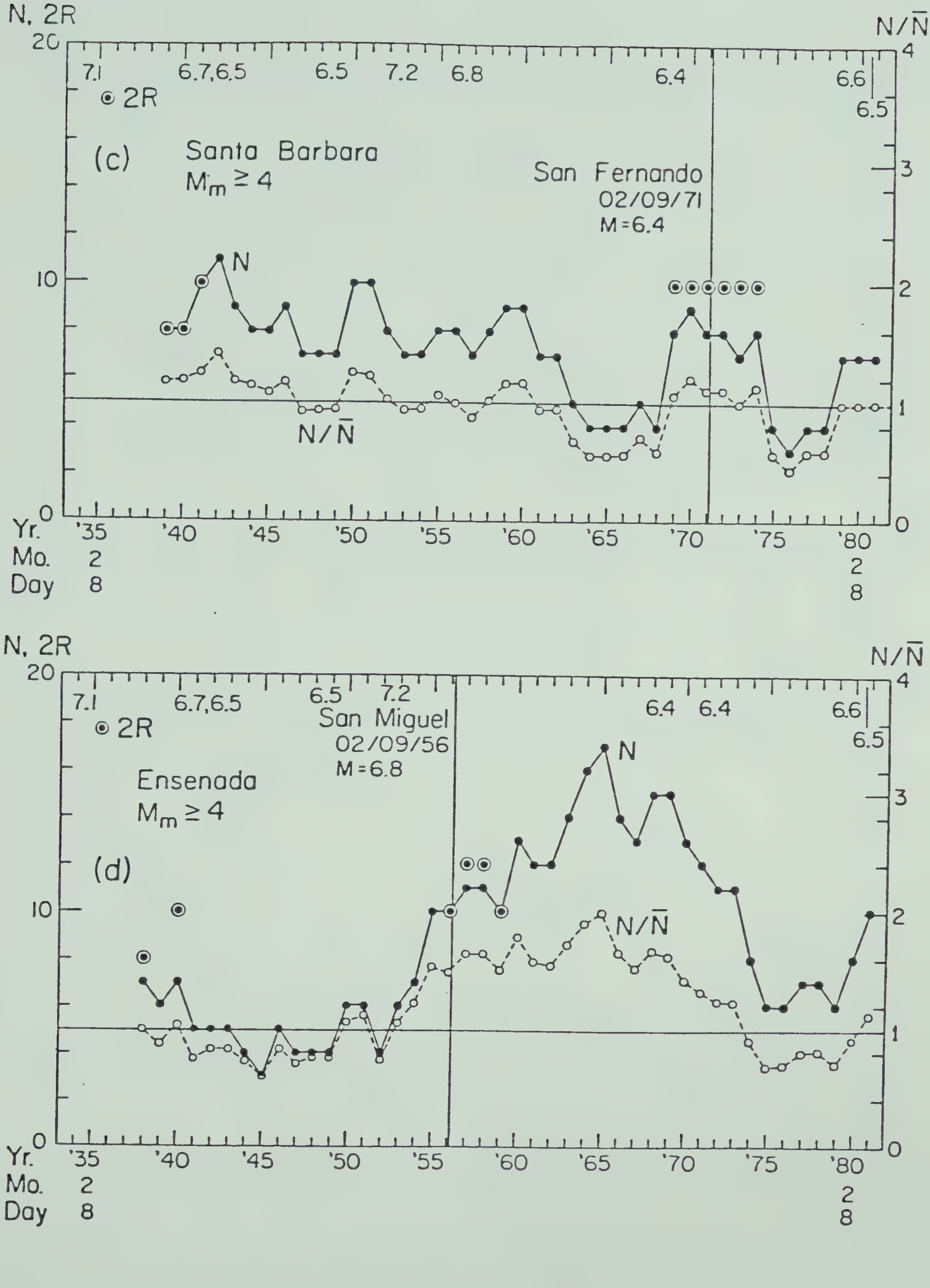
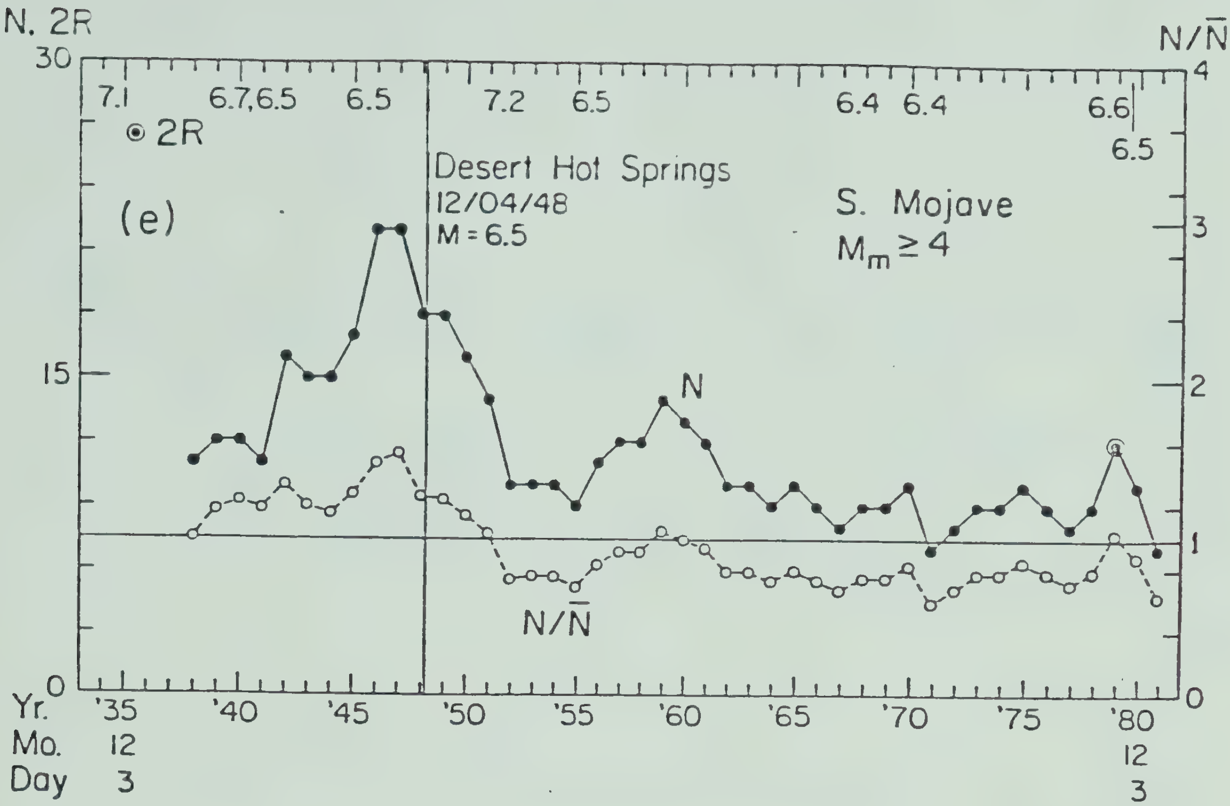


Figure 3.4 gives the test of the quiescence → activation hypothesis for the Santa Barbara and Ensenada regions.



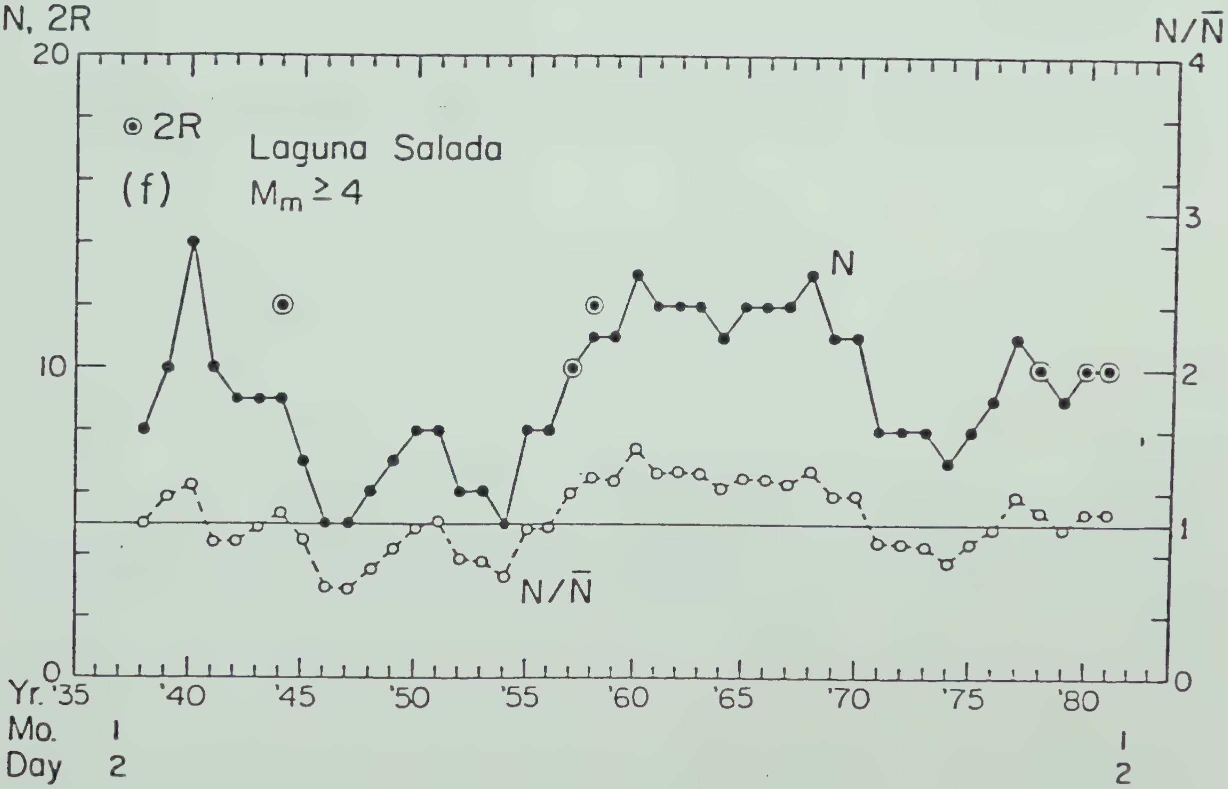


Figure 3.5 gives the test of the quiescence → activation hypothesis for the Southern Mojave and Laguna Salada regions.

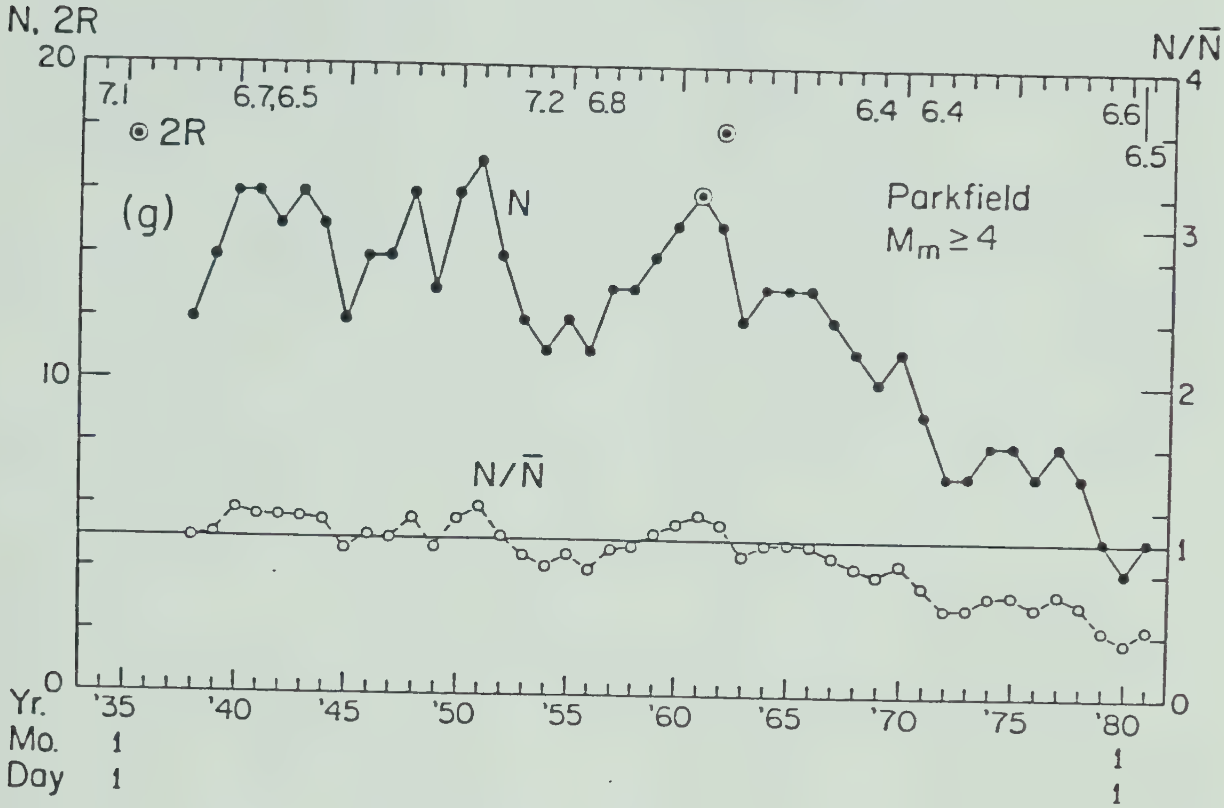


Figure 3.6 gives the test of the quiescence → activation hypothesis for the Parkfield region.

area is presently in a state of quiescence. For the Laguna Salada area a "false alarm" occurred in 1957. This "false alarm" may be connected with a general quiescence - activation associated with the neighboring Ensenada region before the San Miguel earthquake of 1956.

The second group includes the 7 remaining areas where mainshocks with magnitudes less than 4 were considered (figures 3.7 to 3.10). The expected pattern was not observed in these areas. Stronger mainshocks, Borrego Mountain, 1968, and Kern County, 1952, were preceded by swarms, but not by clear quiescence. The same was true for the strongest mainshock in the Southern Sierra region, M = 6.3. In the Riverside area we have a "false alarm" and in the Elsinore area - two "false alarms". In spite of the general poor score for these regions, the recent swarm in the San Jacinto area deserves attention, since it was preceded by a clear minimum of $N(t)$ and $N(t)/\bar{N}(t)$. This swarm includes a mainshock in 1981. The adjacent Southern Mojave area shows a similar pattern. The two areas, San Jacinto and the Southern Mojave, are on opposite sides of the present seismic gap identified in Thatcher et al. (1975). This gap is located on the San Jacinto fault in the northwest corner of the San Jacinto region (see figure 3.1).

In both groups the score could probably be improved by the change of parameters, first of all by the use of the thresholds from Table 2.1 and by narrowing Δx and Δy in the definition of a swarm. However, as a posteriori improvement,

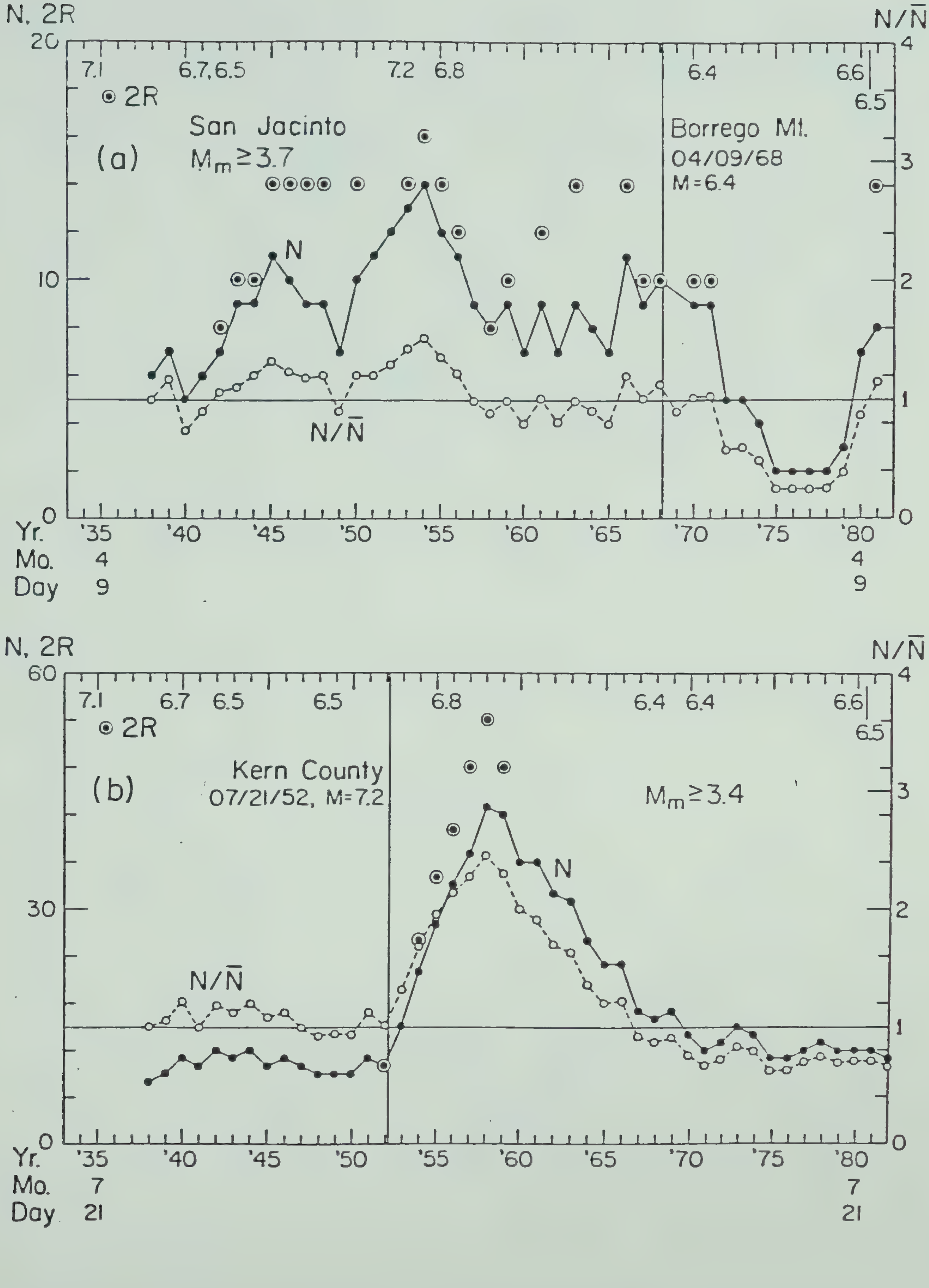


Figure 3.7 gives the test of the quiescence → activation hypothesis for the San Jacinto and Kern County regions.

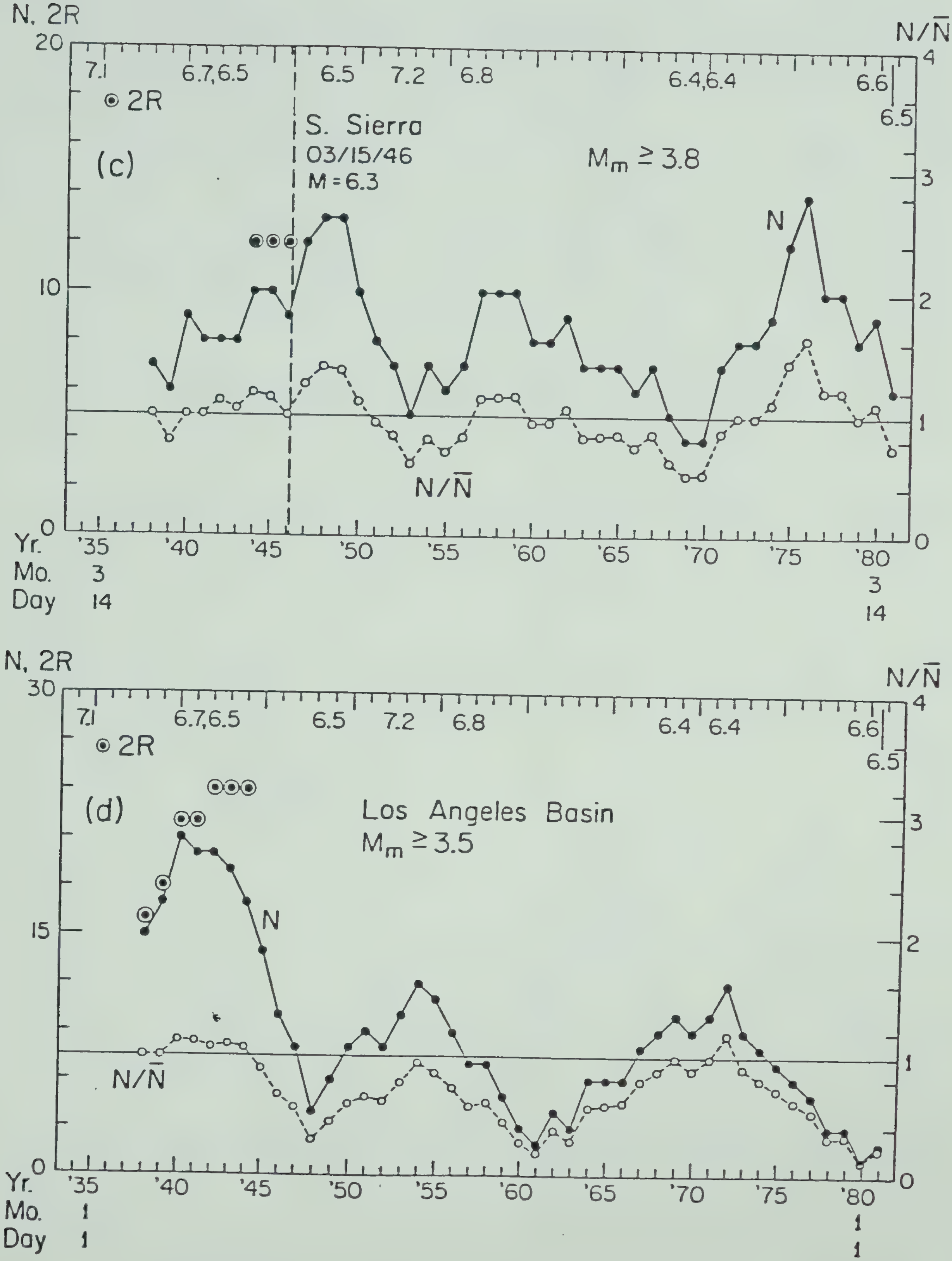


Figure 3.8 gives the test of the quiescence → activation hypothesis for the Southern Sierra and Los Angeles Basin regions.

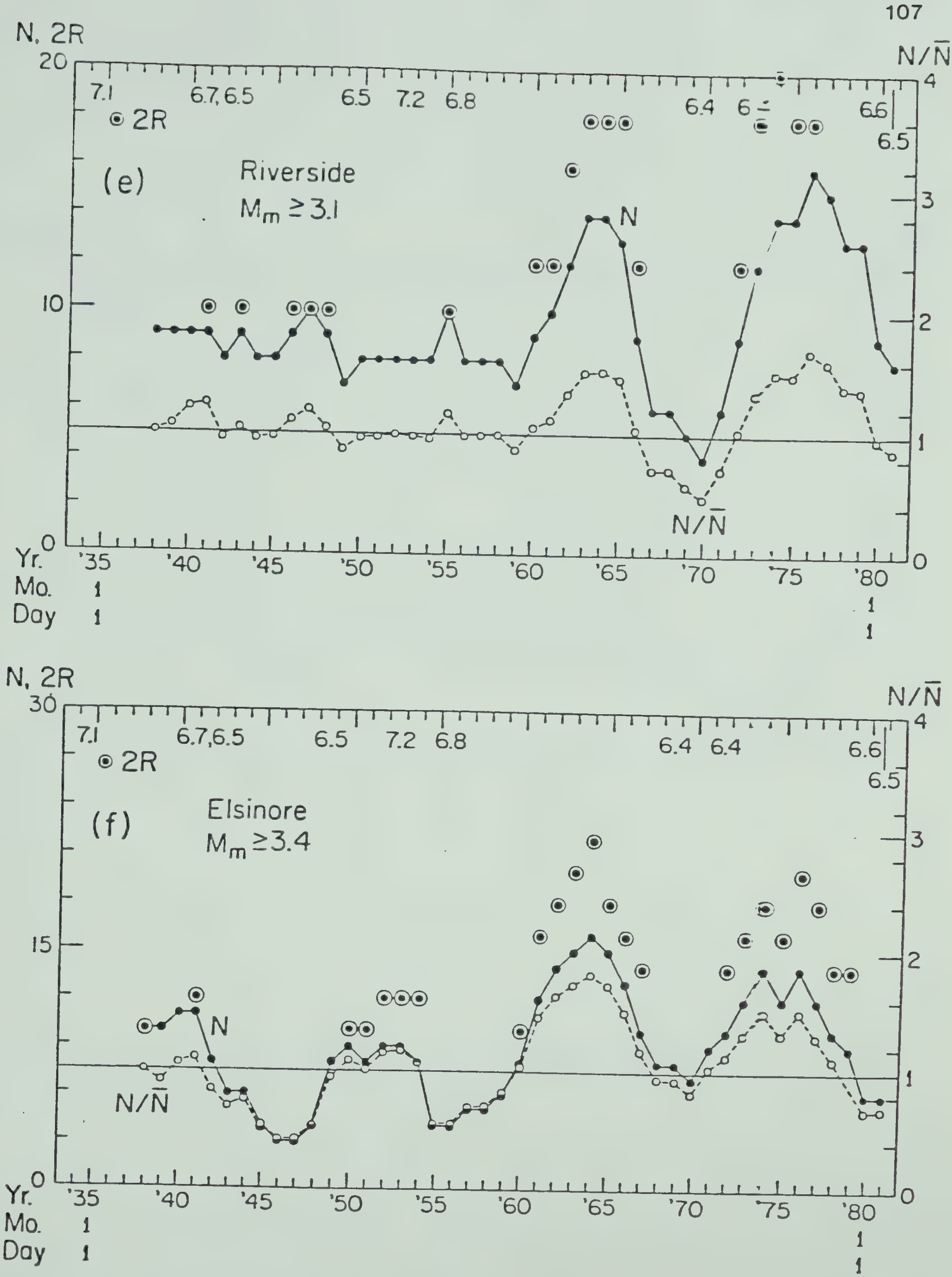


Figure 3.9 gives the test of the quiescence → activation hypothesis for the Riverside and Elsinore regions.

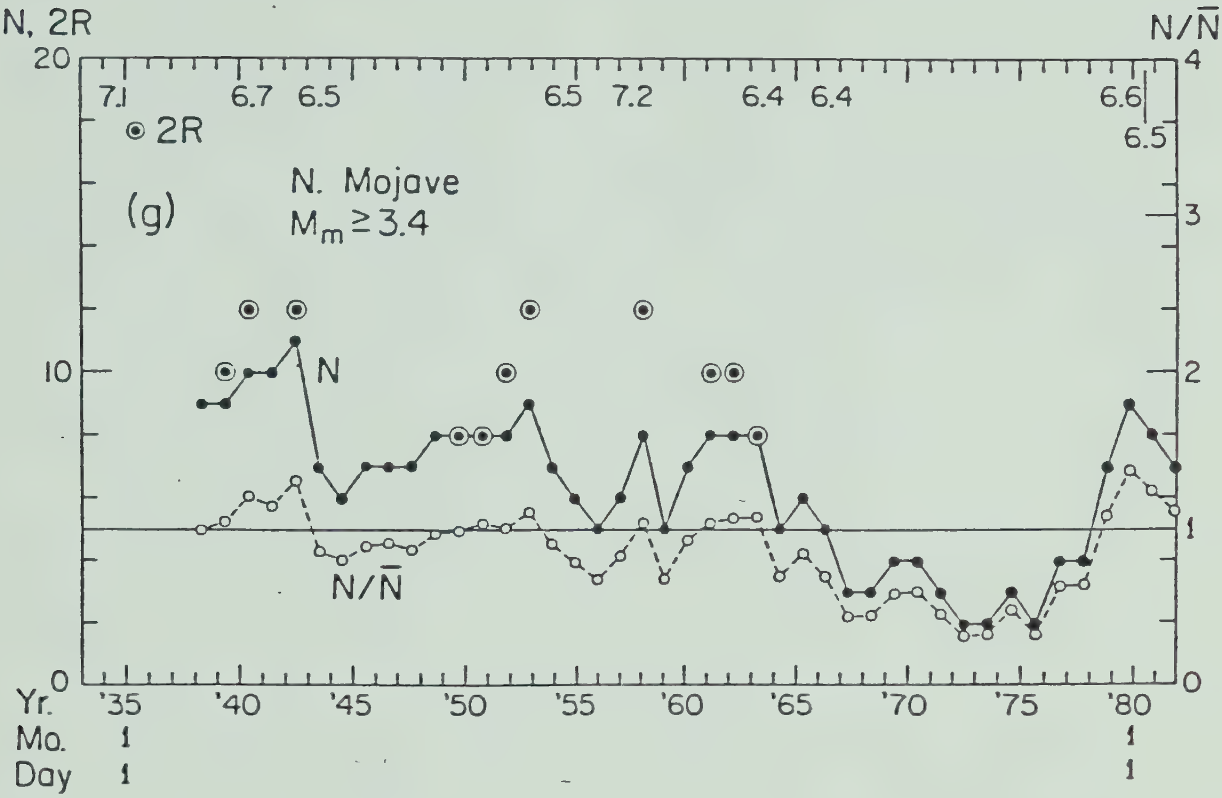


Figure 3.10 gives the test of the quiescence → activation hypothesis for the Northern Mojave region.

it would hardly change the conclusions. Figure 3.11 gives a rough qualitative summary of the results for all areas.

3.6.5 Conclusions: part 1

Mainshocks of Southern California, while forming a Poisson sequence in this region as a whole, are sometimes clustered in the space-time domain. This clustering will be more thoroughly examined in the following sections of this chapter were a more general definition of clusters is used.

There are qualitative confirmations of the hypothesis that mainshocks of magnitude $Mm \geq 4$ form a "quiescence → swarm" pattern before a mainshocks of $Mm \geq 6.4$ in the same fault zone. For a more qualitative test of the hypothesis the follwing are required: quantitative definitions of quiescence; more objective separation of the fault zones, or a method to identify quiescence and clustering without such separation; and a statistical model of the sequence of mainshocks.

3.7 Part 2: cluster analysis of mainshocks in Southern California

The first part of this chapter demonstrated that the sequence of mainshocks in Southern California, as defined by the simple criteria of Table 2.1, shows a significant degree of local clustering. Periods of low clustering (quiescence) followed by high clustering (swarms) were suggested to be

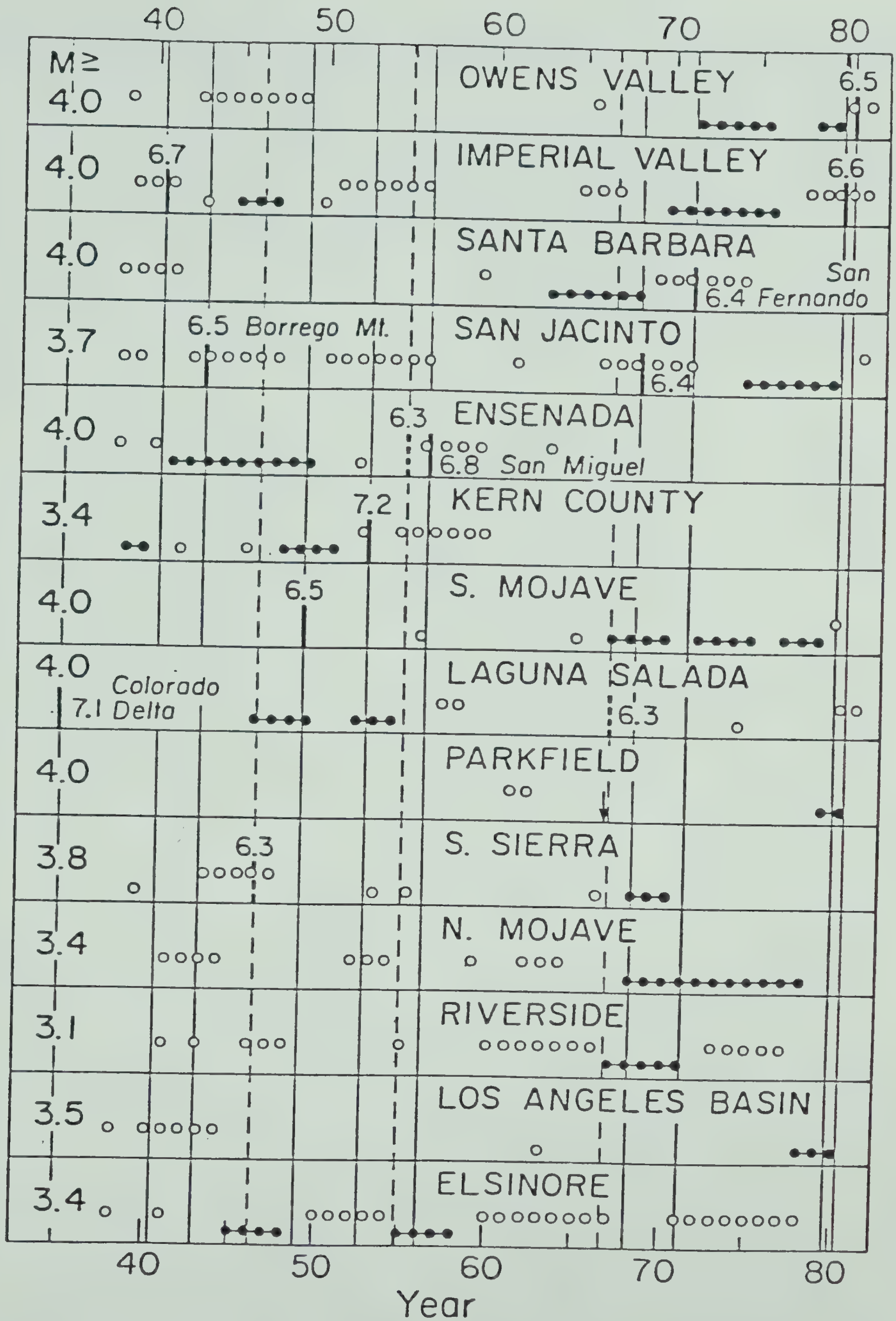


Figure 3.11 gives a rough summary of the results of of part 1. Periods of quiescence is indicated by dark circles and swarms are indicated by open circles. The minimal magnitude for the region and large earthquakes occurring in the region are also indicated.

premonitory to a future strong earthquake near the quiescent swarm region. The swarm model of clustering was very simplistic and the characteristics of individual swarms were not considered. This section uses cluster analysis to investigate the clustering of the mainshocks on a more formal level. The sequences of clusters in several areas are examined.

Cluster analysis is a means of classifying a group of individuals (objects, cases, specimens etc.) into sub-groups such that the "similarity" between individuals within each sub-group is high while the "similarity" between individuals of different sub-groups is low. Although the application of cluster analysis is most familiar in the social sciences, the need for classification schemes is common to all fields of study. With an adequate classification scheme, the amount of data necessary to describe a population can be greatly reduced (i.e. the population can be characterized by the characteristics of the sub-groups rather than the characteristics of all members of the population). Remote sensing is an example of an area in the geophysical sciences which has made extensive use of cluster analysis (for example: Holmes and MacDonald, 1969; Landgrebe et al., 1969; and Landgrebe and Phillips, 1967). The classification problems encountered in pattern recognition and those in cluster analysis are very similar.

The amount of literature on cluster analysis is large. The nature and technique of applications of cluster analysis

vary significantly from problem to problem. Only the concepts of cluster analysis which are used in this chapter will be discussed. For a comprehensive review, the interested reader is refered to to the following sources: Wishart (1978); Jardine and Sibson (1971); Duran and Odell (1974); and Spath, (1980).

3.7.1 The clustering problem and mainshocks

Let $\bar{Q} = \{q_1, q_2, q_3, ...\}$ denote the set of N mainshocks in Southern California. This set forms the population for the study; the individual mainshocks are the elements. Each earthquake is characterized by a set of observables $\bar{X}$ which are used to define the similarity or dissimilarity between elements. In theory these observables could include any characterisitics of the mainshocks such as location (latitude, longitude), focal depth, time of occurrence, focal mechanism, magnitude etc. Only latitude, longitude, and time of occurrence will be used in this study. Let J be an integer less than N. The problem posed by cluster analysis is then to determine J subsets (clusters) of $\bar{Q}$ such that each element is assigned to one and only one cluster and such that individuals assigned to the same group are "similar" while individuals in different groups are "different", the measure of similarity and dissimilarity being determined by the set of observables $\bar{X}$. One group may be a residual set which contains the individuals that are not similar to other groups of individuals.

The terms similarity and dissimilarity need to be quantitatively defined. Needed is a measure of the similarity or dissimilarity between two elements, an element and a cluster, and between two clusters. Many such measures have been proposed (see for example Wishart, 1978). The most familiar are measures of dissimilarity based on the ordinary Euclidian distance which are used in this study.

3.7.2 On Uniqueness

Any classification scheme devised is inherently non-unique. The most obvious source of non-uniqueness enters in J the number of clusters determined. Obviously, a good solution depends on a reasonable choice of J and even for a given J there is no guarantee of uniqueness. Some criterion is needed to measure the "goodness" of the solution. If such a criteron can be properly constructed, the partitioning can be formulated as an optimization problem based on the desired criterion.

Even when such a problem can be formulated, the problem of non-uniqueness still exists and different "goodness" criteria could give different results. The problem of the relative scaling of the various observables is another source of non-uniqueness. For example, when time and spatial coordinates are included, the relative weights of the two must be defined in the dissimilarity criteria (some unit of time must be equated to some unit of distance). In general, the clusters will vary under different scales. Also, some

computational algorithm is needed to solve the clustering problem. There is no guarantee that a computed solution will give an absolute optimum and not just a local optimum.

A direct way to solve the clustering problem for a given optimization criterion is to evaluate the optimization criteria for each choice of clustering alternative. The unique solution would then be the choice yielding the optimal value. However, this method, termed clustering by complete enumeration (Duran and Odell, 1974) is impractical for any reasonable problem. The number of possible partitions of a set of n objects into j subsets none of which is empty is given by S(n,j), where S(n,j) is the Stirling number of the second kind (Abramowitz and Stegun, 1968).

For a six year interval there are on the average 200 mainshocks with magnitudes greater than 3.5 in Southern California. If ten clusters are desired, complete enumeration would require the evaluation of approximately 10^{194} cases. Some efficient strategy is needed to limit the amount of computational time required to some feasible value.

Many efficient clustering algorithms have been suggested in the literature. Three different strategies are used in this study, density analysis, hierarchical fusion, and iterative relocation. These algorithms are taken from CLUSTAN, a cluster analysis library published by Edinburgh University (Wishart, 1978).

3.7.3 Notation and definitions

The following definitions and notations from Wishart (1978) are needed:

-- $\bar{N}$ .... the total number of individuals.

-- m .... the number of variables per observation.

-- J .... the total number of clusters.

-- $\bar{X}i$.... vector of length m giving the observations for the ith individual.

-- Xij .... the value of the ith variable for the jth individual.

-- Ni .... the number of individuals in the ith cluster.

-- Zik .... the mean value of the ith variable for the kth cluster.

-- $\bar{Z}i$ .... a vector of length m giving the center of the ith cluster.

-- Dij .... the ordinary Euclidian distance between the centers of the ith and jth cluster.

$$D_{ij}=\frac{1}{m}\left(\sum_{l}\left(Z_{li}-Z_{lj}\right)^{2}\right)^{1/2} \qquad 14$$

-- Err(i) ... the error sum of squares for the ith cluster.

$$Err(i)=\frac{1}{m}\left(\sum_{k\varepsilon i}\sum_{n}\left(X_{nk}-Z_{ni}\right)^{2}\right) \qquad 15$$

-- I(i,j) ... the increase in the error sum of squares

when clusters i and j are joined (i.e. $I(i,j) = Err(i+j) - Err(i) - Err(j)$)

The error sum of squares, refered to as the within-group sum of squares by Duran and Odell (1974), is the sum of the distances from the center of a cluster to each individual contained in the cluster. As such it is a measure of the scatter around the cluster center. Two of the cluster techiques used here (hierarchical fusion and iterative relocation) use this functional as the optimization criterion. Solutions are sought which minimize the total error sum of squares for J clusters.

3.7.4 Algorithms used

3.7.4.1 Hierarchical fusion

Methods of hierarchical fusion start with N clusters (each individual is a starting cluster). In each of N-1 cycles the two "closest" clusters are fused forming a single cluster. The final step consists of a single cluster of size N. The method can be thought of as giving the sequence of overlapping classifications which define the hierarchical manner in which each individual and/or group of individuals joins the entire population. The method of hierarchical fusion used here was first suggested by Ward (1963). In Ward's method each cycle combines the two clusters whose fusion yields the least increase in the total error sum of squares. Ward's method finds minimum variance spherical

clusters (Wishart, 1978).

3.7.4.2 Iterative relocation

Iterative relocation finds a local optimum for J clusters, starting with an initial classification of the individuals into J initial clusters. Each individual is removed from its parent cluster. The similarity between the individual and each cluster is compared and the individual is placed (relocated) in the cluster which results in the least total error sum of squares. The process is repeated until there are no relocations during one complete cycle (a scan of each individual). In general the process can be shown to converge for reasonable data. However, convergence will not in general find a global optimum in the error sum of squares for the J clusters. Rather, convergence will be towards a local optimum in the neighborhood of the initial cluster classification (Wishart, 1978). A useful grouping by iterative relocation depends on a good starting classification which is not too distant from the final desired solution. In this study iterative relocation will be used to refine initial classifications generated by other methods.

3.7.4.3 Number of natural clusters: density analysis

Suppose $\partial(x,y,t)$ is some continuous function which gives a theoretical model of the density of epicenters in space and time. Wishart (1969) suggests that the number of clusters for the lowest level of clustering at which the

classifications are "natural" or taxonomically significant is given by the number of local maxima in such a density function. The lowest level of classification corresponds to the hierarchical level with the maximum number of clusters. This will be taken as the number of clusters of interest in this study.

To determine the number of maxima it is necessary to construct an estimate for the continuous density function using the discrete set of epicentral locations given by the data. Some averaging process is necessary. One method is to divide the space into small cells in latitude, longitude and time. An estimate of the density at a location could then be given in terms of the number of events per cell. This is not a good method since local maxima which are close could be fused in the averaging process. Wishart (1978) suggests that the density estimate at a given location be based on the average distances to some number of nearest events.

A pseudo density $\hat{\partial}(x,y,t)$ is defined here as follows: let $\{D_{i1}, D_{i2}, \ldots D_{ik}\}$ be the distances from the ith event to its k nearest neighbors. Then $\hat{\partial}(x,y,t)$ is given by

$$\partial(\bar{X}_n) = k / \sum_{m=1}^{k} D_{nm} \qquad 16$$

The number of nearest neighbors k used in the above estimate determines the amount of averaging used in the calculation of $\hat{\partial}$. An increase in k is analogous to

increasing the volume of a cell over which average densities are estimated. Two local maxima which are close will remain distinct if each is a result of more than k/2 events.

Physically $\hat{d}$ gives an estimate of the density in the neighborhood of an individual by using the estimated radius which encloses k/2 neighbors, this radius being estimated as the average distance to the k nearest neighbors. Wishart (1978) points out that this procedure will give smoothed density estimates. The densities at the maxima will be underestimated and the densities at saddle points will be overestimated. Maxima due to less than k/2 individuals will not be detected.

3.7.5 Initial clustering constraints

In order to search for clusters of mainshocks some preconception of what constitutes a cluster is necessary. The results of part 1 of this chapter will be used for this purpose. In the first part of this chapter swarms were defined as groups of mainshocks occurring within 6 years or less of each other, and occurring within a rectangle of .4° by .4°. The results of Section 3.4 can be used suggest more appropriate time and distance scale for the clusters.

Section 3.4 suggests that "mainshocks" (as defined here) with magnitudes greater than 4.0 show a significant degree of clustering for time periods of the order of 3 to 6 years and distances of the order of 100 km. Somewhat shorter time and distance scales were suggested for the case with

minimal magnitude equal to 3.5. The possibility of even longer term clustering in the catalogue could not be tested. The question asked here is, what sequences of earthquakes give rise to clustering characteristics for these time and distance scales? Although this observation is not well constrained and does not give a optimal estimate of possible spatial and temporal sizes for the clusters, it can be used to give an order of magnitude estimate of these sizes. In particular, the results of Section 3.4 give an order of magnitude estimate for the scaling of time relative to distance. This suggested normalization, to look for clustering as was indicated in Section 3.4, is: 3 to 6 years equal to 100km.

3.7.5.1 Dissimilarity criteria

In terms of latitude (y), longitude (x), and time Dij is given as

$$D_{ij}=\frac{1}{3}\left\{\frac{(x_i-x_j)^2}{x_0^2}+\frac{(y_i-y_j)^2}{y_0^2}+\frac{(t_i-t_j)^2}{t_0^2}\right\}^{1/2} \qquad 17$$

where x_0, y_0 and t_0 are normalization factors. The constants x_0, y_0, and t_0 are chosen so that at the center of the region (35° latitude) a specific unit of distance, r_0, is equal to a specific unit of time. Examples used here include: 50 km. = 2 years; 50 km. = 3 years; 75 km. = 2 years; and 30 km. = 2 years. These result in normalizations of 100 km. = 6.67, 6.0, 4.0 and 2.57 years. There is no physical basis for these choses other than the observations

of Section 3.4.

As an example x_0=.548393°, y_0=.449507°, and t_0=2 years give the following properties to Dij:

-- At the center of the region (35° latitude) two years of time is equated to 50 kilometers of spatial distance.

-- Two events occurring at the same place and separated by 6 years will give a Dij of 1.

-- At the center of the region two events occurring at the same time and separated by 150km will give a Dij of 1.

3.7.5.2 Residual events

The results of Section 3.4 suggest that clustering of mainshocks, although present, may be weak. Some earthquakes may be far removed from areas of clustering. Local clusters are sought. These local clusters need not include all events. For this reason a residual set is created which includes those events not found to be sufficiently similar to the clusters found.

For practical purposes clusters with very few individuals are not of interest. Only clusters with more than five individuals were considered. This minimum number was chosen arbitrarily.

To create the residual set the following constraints were applied to the events:

1. If a cluster was found to contain less than five individuals the individuals were placed in the residual set if the individuals were not sufficiently close to another

cluster.

2. Let k be the number of the cluster to which the ith event is assigned. If Dik > $\bar{T}$ the event is moved to the residual. Dik is the dissimilarity between the ith event and centroid of the kth cluster with the ith event removed and $\bar{T}$ is some threshold value.

Events which are in the residual set do not necessarily remain in the residual set. If the centroid of the clusters changes so that the minimum dissimilarity between an event and the closest cluster is less than $\bar{T}$, the event is assigned to that cluster. At the end of the clustering process the residual set is searched for the existence of missed clusters. The value of $\bar{T}$ limits the time and spatial distance that a cluster can span. For example, a choice of $\bar{T}$ = 1.00 limits the time span of a cluster to less than $3t_0$ years and the spatial span of a cluster to less than $3r_0$ at the center of the region. $\bar{T}$ will equal 1.00 in this study.

3.7.5.3 General clustering procedure

The procedure used to find clusters is as follows. A density function $\hat{d}$ was constructed as indicated in section 3.6.4.3 to estimate the maximum number of clusters for the analysis. Several values of k were used to construct $\hat{d}$. The results presented here are based on k=7 for M≥4.0 and k=11 for M≥3.5. The number of local maxima (J) in the density function were determined. This number was equated to the maximum number of clusters searched for in the study. In

general the number of clusters that resulted was less than J. Several of the local maxima were found in areas of low density and did not survive the threshold criteria.

Two separate procedures were used to generate initial classification of all objects into J initial clusters. First, events were grouped into J clusters by assigning each event to a cluster that had an assumed centroid corresponding to the location of the local maximum (i.e. each event was assigned to the nearest local maximum). Second, Ward's method of hierarchical fusion was used to generate J+I cluster categories. Ward's method could result in the splitting of a cluster due to just one local maximum into several clusters. "I" was chosen by looking at all hierarchical classifications with J or more clusters and picking as the desired initial classification the hierarchical level with the minimum number of clusters which had at least one cluster for each dense point. In both methods, all events are classified into clusters (no threshold criterion was used).

Iterative relocation was then used to refine the two initial classifications by creating a residual set and the final results were compared. The results of the separate procedures differed very little. The results using hierarchical fusion to generate initial classifications are presented here. In general they resulted in final classifications with lower total error sum of squares than did the other procedure. The entire process was repeated on

the residual set to search for missed clusters.

3.7.6 Results

The output of cluster analysis consists of a listing of the J groupings of the earthquakes, each grouping constitutes a single cluster. For Southern California such listings are rather large, consisting of 100 to 300 clusters (depending on the clustering parameters) of the approximately 2000 mainshocks with $M \geq 3.5$. A complete listing of cluster groupings will not be given here. Rather, several illustrative examples will be discussed. A listing of several complete cluster analysis groupings in computer readable form is provided for the interested reader on tape number 007917, volume label LMRX10 stored at the University of Alberta Computer Centre. A copy of the tape is available upon request from Dr. Edo Nyland, Department of Physis, University of Alberta. The tape consists of 9 separate files. File 1 contains a complete description of the contents and formats of all other files. Listings and explanations of several Fortran programs needed for the interpretation of cluster analysis results are also contained on the tape along with the following cluster analysis examples:

1. Cluster analysis for the northern section of Southern California (see Figure 3.12 for region boundaries) with $M \geq 4.0$, $k=7$, and a normalization of 50km=3years.
2. Analysis for the southern region of Southern California

with $M \geq 4.0$, $k=7$, and a normalization of 30km=2years.

3. Analysis of the northern section of Southern California with $M \geq 3.5$, $k=11$, and a normalization of 50km=2years.

4. Analysis of the southern section of Southern California with $M \geq 3.5$, $k=11$, and a normalization of 30km=2yrs.

The interested reader is invited to analyze these examples or to generate other classifications using the procedure outlined on the tape.

Two examples of generated clusters are given in Table 3.9 and 3.10 and Figure 3.12. Due to computing cost considerations it was necessary to divide the Southern California region into two segments for the cluster analysis. The boundaries of these segments are given on figure 3.12. Table 3.9 gives a listing of the centroids of clusters found in the Northern section with normalization factors of $x_0=.548393$, $y_0=.449507$, and $t_0=3$ years (i.e. 50 km. equal to 3 years). Also given are the standard deviations in latitude, longitude and time for each cluster and the total Sigma of each cluster. Sigma is defined as

$$\mathcal{S}_i = \sum_{k \varepsilon i} 10^{.91(M_k - 4.5)} \qquad 18$$

Sigma is roughly proportional to a summation of the fracture areas of the earthquake sources (Keilis-Borok and Malinovskaya, 1964).

Table 3.10 gives a similar listing for clusters in the southern section with $x_0=.329036$, $y_0=.269704$,and $t_0=2$ years (i.e. 30 km. equal to 2 years). Figure 3.12 shows the

n	lat.	d(lat.)	long.	d(long.)	time	d(time)	Sigma
		centroid for all events					
290	36.15	1.4825	-118.71	0.8024	1954.35	13.8063	
		centroid for the residual events					
52	35.70	1.5295	-118.90	1.1945	1951.99	15.0746	
		centroids of the 33 clusters					
5	35.79	0.1372	-118.23	0.1930	1934.84	1.9505	6.851
13	37.96	0.2976	-118.23	0.2358	1936.37	1.9615	40.789
5	35.10	0.3191	-119.02	0.1549	1935.51	0.8054	5.000
8	34.47	0.1191	-120.62	0.1669	1937.78	1.8129	13.553
6	35.98	0.2702	-117.50	0.4279	1939.13	0.8896	7.851
7	34.38	0.2403	-119.87	0.3589	1942.95	2.5907	80.409
10	37.54	0.0495	-118.68	0.0933	1941.74	1.5678	93.736
7	34.90	0.3585	-119.05	0.1598	1941.38	1.3806	22.061
13	35.97	0.2567	-117.99	0.1130	1944.93	1.9774	31.096
10	37.32	0.1849	-118.43	0.2502	1945.77	1.2978	17.754
5	37.15	0.1026	-117.51	0.3312	1948.21	2.2703	42.309
7	34.91	0.3637	-119.00	0.1296	1949.35	2.2641	11.401
5	33.68	0.3850	-119.27	0.1961	1949.30	1.9843	16.859
8	37.66	0.2242	-118.58	0.2394	1950.50	1.2940	25.067
11	35.37	0.1589	-118.50	0.1947	1953.35	1.2724	31.498
6	37.62	0.1133	-118.52	0.2356	1954.67	0.9762	10.020
7	34.09	0.2749	-119.40	0.2815	1956.06	1.5947	18.769
7	34.83	0.2953	-118.91	0.3126	1956.43	1.0337	12.977
8	38.24	0.1996	-119.05	0.2573	1958.07	1.6809	13.928
5	35.83	0.1845	-117.91	0.2234	1958.72	1.8708	23.617
7	37.54	0.2243	-118.51	0.2395	1959.57	0.9535	47.528
6	35.10	0.2060	-118.89	0.2403	1962.33	1.7711	23.592
7	38.33	0.2253	-118.26	0.2813	1964.45	2.1657	18.819
5	35.48	0.2726	-117.80	0.2845	1964.32	2.2291	18.370
5	38.31	0.0642	-119.26	0.1101	1963.93	1.4173	29.785
7	37.49	0.0816	-118.56	0.4119	1964.82	1.6270	15.815
5	35.73	0.3353	-118.43	0.0952	1971.16	2.5796	5.233
5	34.22	0.0600	-119.66	0.0554	1968.50	0.0082	18.425
6	35.55	0.2212	-117.56	0.1858	1972.97	1.9467	8.586
6	37.67	0.1591	-118.52	0.3706	1973.43	1.6869	8.749
5	35.02	0.1618	-118.95	0.2404	1974.72	2.3361	9.913
7	34.14	0.2283	-118.73	0.2692	1976.91	2.4975	74.930
14	37.69	0.2368	-118.74	0.2337	1980.23	0.5332	104.734

Table 3.9 gives a tabulation of the centroids for clusters found in the northern region of Southern California for the case $M_0=4.0$ and 50 km. equal to 3 years (i.e. $t_0=3$ years, $x_0=.548393°$ and $y_0=.449507°$).

n	lat.	d(lat.)	long.	d(long.)	time	d(time)	Sigma
		centroid for all events					
528	32.72	1.2433	-116.14	1.0142	1955.17	14.2710	
		centroid for the residual events					
98	32.59	1.6645	-116.53	1.4260	1955.49	14.7124	
		centroids for the 52 clusters					
7	30.57	0.5345	-114.00	0.0000	1940.60	0.9675	21.108
5	31.81	0.3633	-116.10	0.2040	1980.40	0.3636	7.857
6	33.55	0.3648	-116.49	0.1788	1933.49	0.6303	10.647
9	32.04	0.2872	-115.12	0.2836	1934.35	0.7566	12.553
10	33.01	0.2403	-115.43	0.2357	1934.96	1.2957	25.928
15	33.83	0.2591	-117.76	0.3150	1935.97	1.4797	48.764
5	32.04	0.0965	-116.62	0.0780	1936.52	1.6998	19.257
11	33.36	0.2015	-116.56	0.3205	1936.97	1.3089	15.553
8	34.32	0.3516	-116.55	0.2772	1937.92	0.9339	20.725
17	32.83	0.3262	-115.85	0.3578	1939.54	0.9704	37.704
8	33.75	0.0621	-118.20	0.1310	1940.24	1.2895	39.906
8	31.52	0.3197	-115.07	0.0441	1940.11	0.7909	19.681
5	31.83	0.2724	-117.50	0.3158	1940.49	1.2409	13.979
5	34.17	0.2514	-117.20	0.2627	1941.13	0.9917	10.186
12	33.97	0.3619	-116.30	0.3079	1942.63	1.1357	68.350
9	33.01	0.2417	-115.95	0.3053	1944.47	1.8113	59.119
6	34.08	0.2157	-116.95	0.2237	1945.00	1.1687	38.364
5	31.18	0.2683	-115.84	0.2191	1945.03	0.2796	23.347
9	31.99	0.2494	-115.25	0.2863	1946.81	1.3901	19.927
9	34.05	0.2102	-116.01	0.3201	1947.36	1.5016	125.631
7	34.87	0.2472	-116.74	0.1570	1949.30	1.1726	13.519
5	30.88	0.1304	-115.14	0.2793	1949.04	0.9207	21.513
14	33.23	0.1518	-116.09	0.4349	1950.57	0.9872	91.975
10	32.09	0.2692	-115.21	0.1572	1953.50	1.6936	75.079
8	34.05	0.2452	-117.16	0.3577	1954.51	1.7067	19.681
10	31.67	0.1965	-116.08	0.1927	1954.62	1.0218	73.839
7	33.31	0.1640	-116.48	0.1578	1954.92	1.5033	12.992
8	32.98	0.2555	-115.78	0.2020	1955.88	1.2178	63.471
8	33.96	0.1530	-116.21	0.2951	1958.10	1.2665	14.852
11	32.37	0.2285	-115.71	0.2538	1959.48	1.4443	71.965
11	31.75	0.2426	-115.39	0.2238	1960.17	1.2967	23.573
9	33.25	0.2642	-116.30	0.2496	1960.77	1.2598	16.821
5	31.82	0.2051	-116.45	0.3054	1960.58	0.9617	13.107
8	30.44	0.4241	-114.25	0.2619	1962.11	1.3644	97.660
8	34.01	0.2169	-116.93	0.4382	1964.18	1.2455	31.797
8	31.81	0.2562	-115.97	0.3788	1963.50	0.8754	30.529
11	31.53	0.2548	-114.58	0.2851	1965.43	1.4737	25.642
9	32.97	0.1890	-115.83	0.2207	1964.25	1.1618	19.677
6	31.17	0.3204	-115.85	0.2510	1965.25	0.9652	21.822
7	33.47	0.3693	-116.26	0.1832	1967.96	1.0905	57.007
8	31.85	0.3134	-116.16	0.3837	1967.55	0.7218	25.666
7	34.07	0.3090	-117.55	0.3252	1969.68	1.4313	28.943
6	31.54	0.2221	-115.55	0.3892	1970.25	1.2981	28.352
7	32.96	0.3088	-116.18	0.2672	1971.54	1.1053	18.555
8	32.30	0.2025	-115.47	0.1632	1973.86	1.4540	17.156
8	34.39	0.1880	-116.55	0.2122	1974.57	1.2748	29.897
9	31.48	0.2878	-115.74	0.2711	1975.08	1.1819	24.948
6	30.45	0.3121	-116.22	0.2656	1975.56	0.6255	34.659
9	33.01	0.1053	-115.63	0.2423	1977.32	1.5600	18.137
6	33.51	0.0977	-116.59	0.1509	1978.25	1.7956	34.450
12	32.12	0.3112	-115.17	0.3481	1978.75	1.1397	37.504
5	34.24	0.0768	-116.87	0.4537	1979.07	0.5648	23.059

Table 3.10 gives a tabulation of the centroids for clusters found in the southern region of Southern California for the case $M \geq 4.0$ and 30 km. equal to 2 years (i.e. $t_0=2$ years, $x_0=.329036°$ and $y_0=.269704°$).

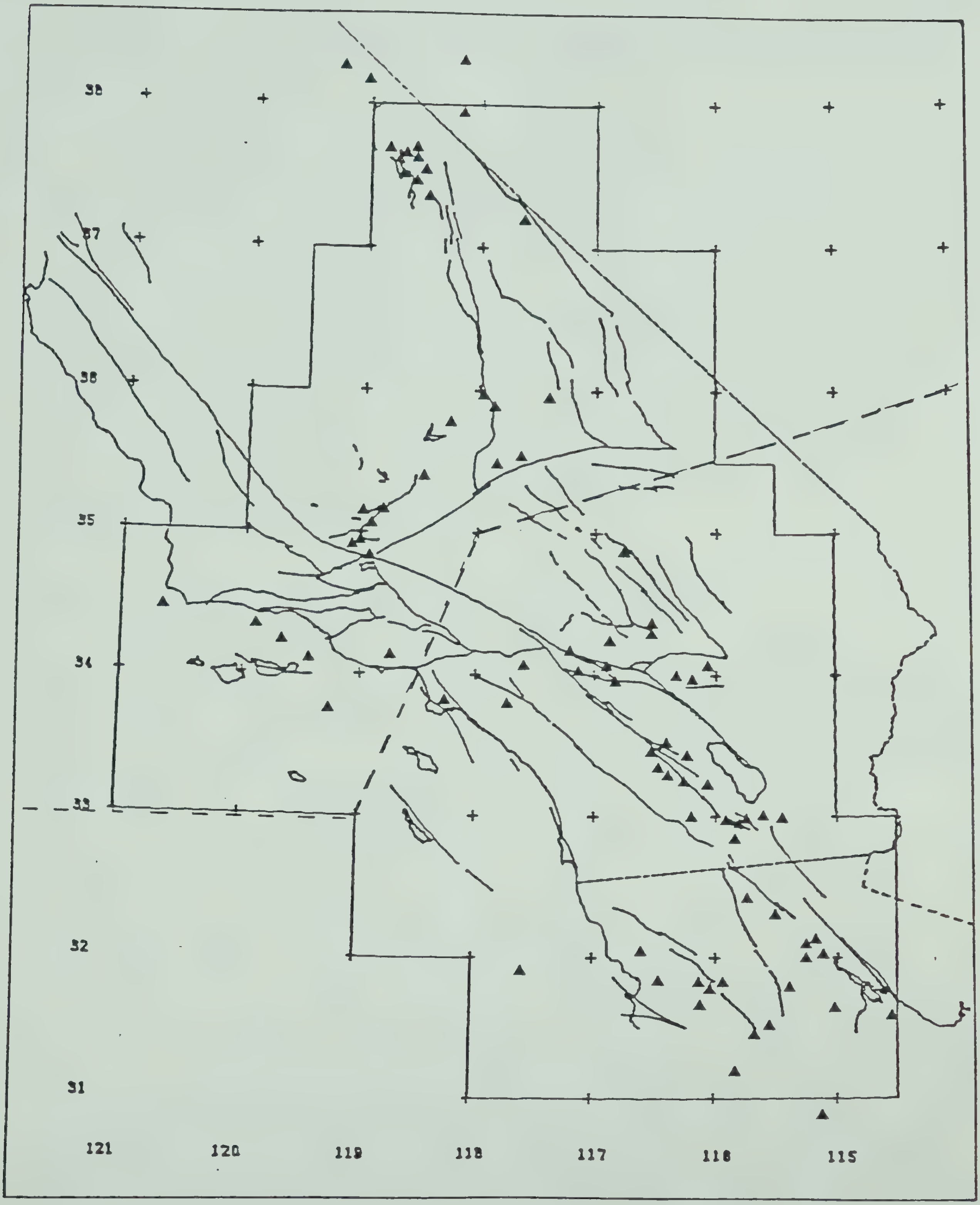


Figure 3.12 gives the locations of the cluster centroids listed in tables 3.9 and 3.10. The boundary between the northern and southern regions is indicated by the dashed line.

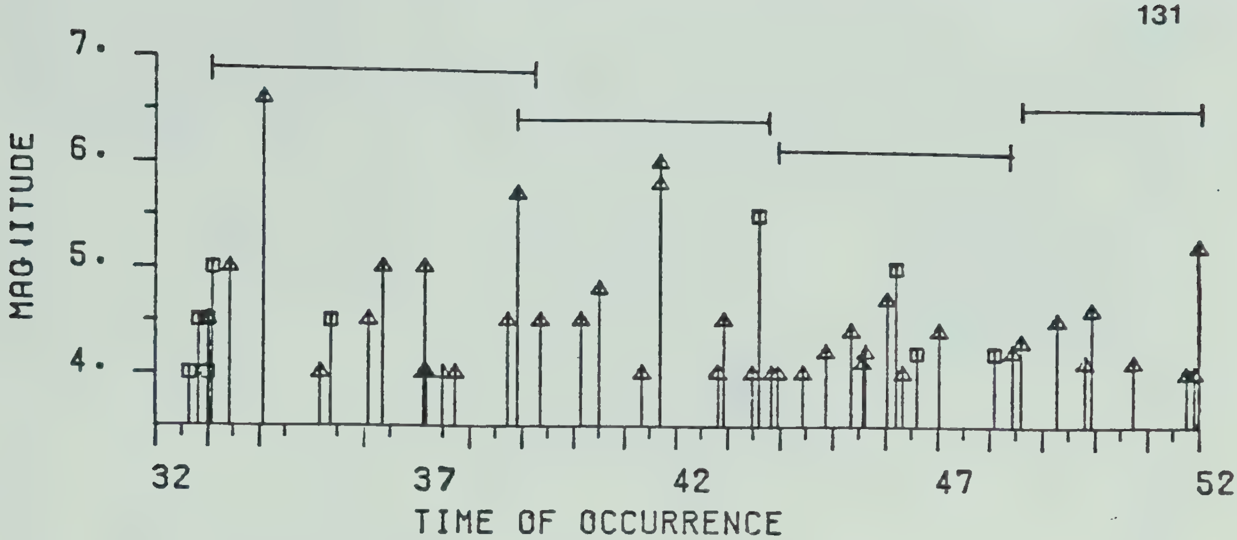
locations of the above cluster centers on a map of Southern California.

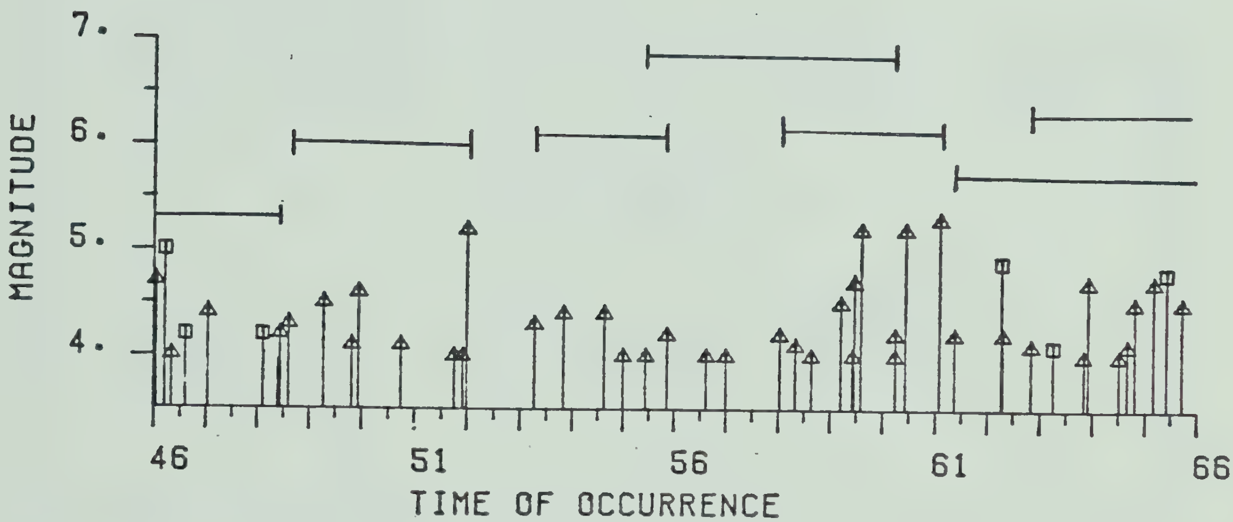
From Figure 3.12 it can be seen that the cluster centers themselves form groups in space. The groups of cluster centers were instumental in the choices of boundaries for the divisions of the Southern California catalogue into the smaller regions analyized in Part 1 of this chapter (Figure 3.1). The remainder of this chapter will give a discussion of the sequence of clusters in several of these subregions.

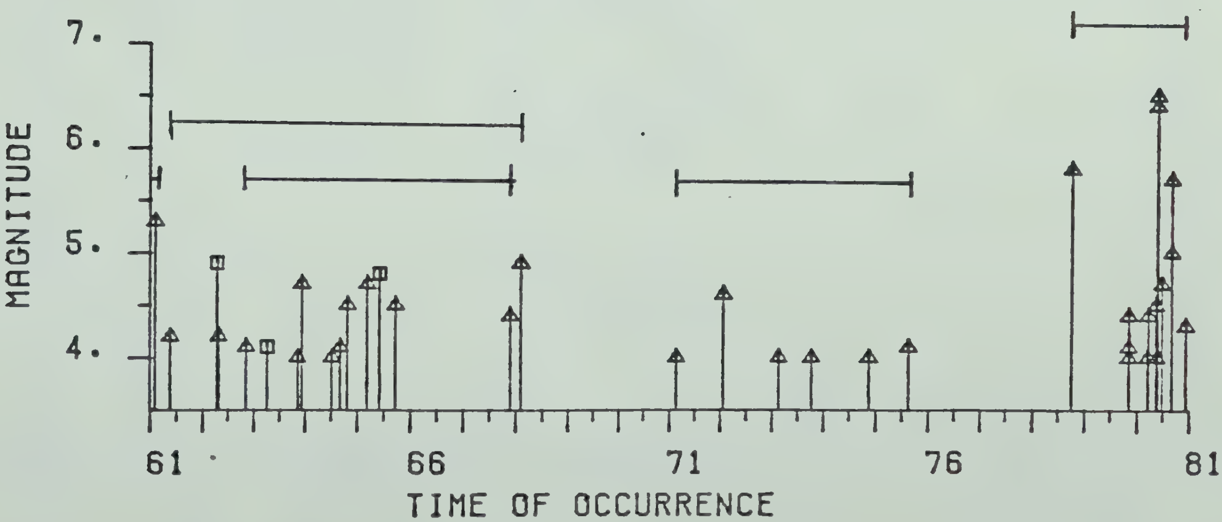
3.7.6.1 Clusters in the Owens Valley region

The Owens Valley region (Figure 3.1) is the most isolated of the regions. Figure 3.13 gives a time magnitude plot of mainshocks, $M \geq 4$, in the Owens Valley region defined by figure 3.1. Earthquakes not found to be members of clusters are shown with a square symbol. The clusters were determined with a normalization 50 km.= 3 years, and K=7. The time spans of the various clusters are indicated by horizontal bars above the plot. Figures 3.14a and 3.14b a latitude, longitude plot for each cluster in the region. The quiescence and activation before the 1980 Owens Valley earthquake is clearly evident in the plots. The quiescence and activation is indicated by number of events, energy release, and compactness of the clusters. The cluster occurring between 1971 and 1976 has the fewest members and lowest Sigma of any cluster in the region. The cluster also

Figure 3.13 gives a time magnitude plot of earthquakes in the Owens Valley region (figure 3.1). Earthquakes not found to be cluster members are indicated bu a square symbol. The times spanned by the clusters are indicated by horizontal bars.


MAGNITUDE
7.
6.
5.
4.
32
37
42
47
52
TIME OF OCCURRENCE


MAGNITUDE
7.
6.
5.
4.
46
51
56
61
66
TIME OF OCCURRENCE


MAGNITUDE
7.
6.
5.
4.
61
66
71
76
81
TIME OF OCCURRENCE

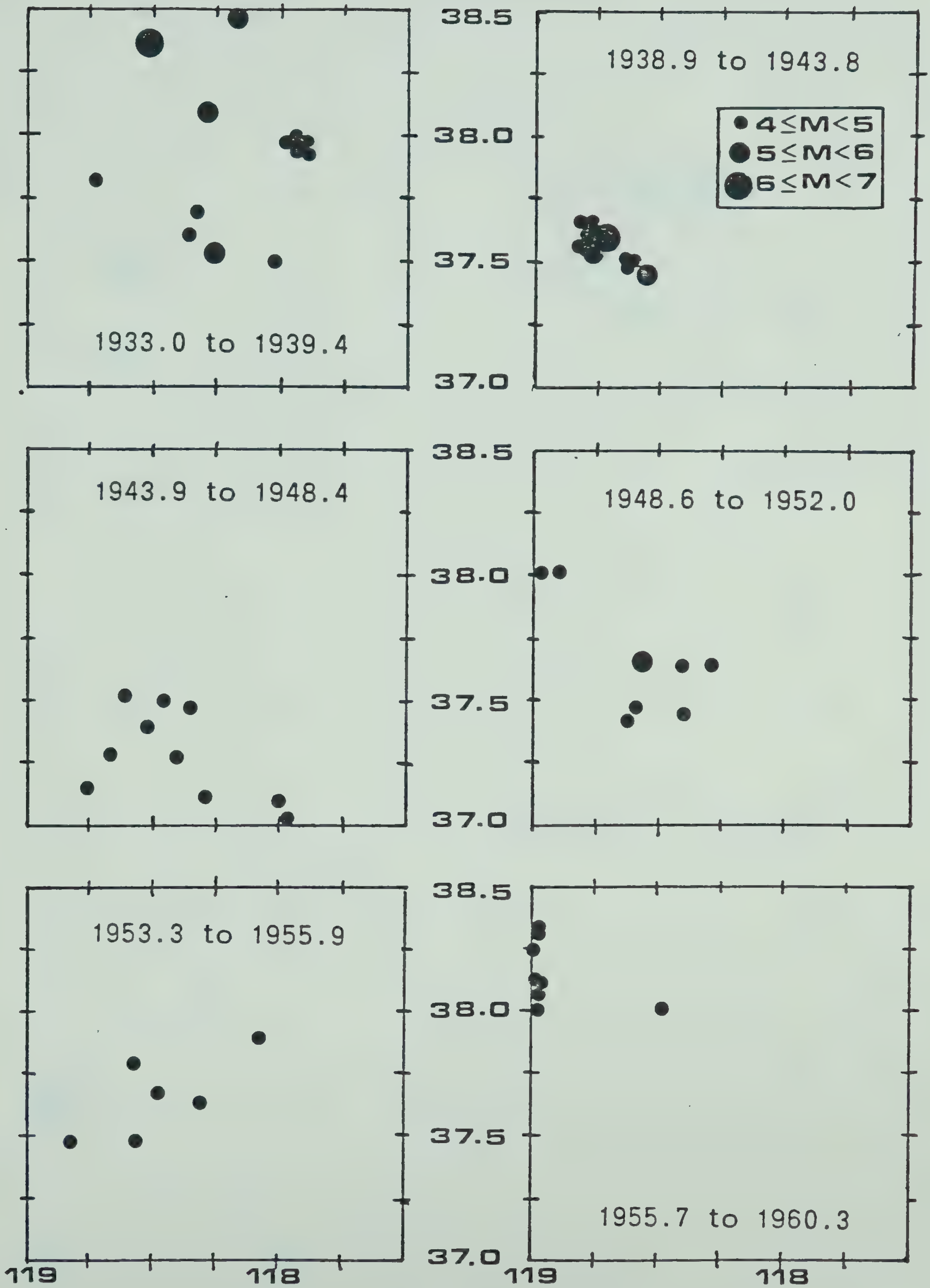


Figure 3.14a gives latitude longitude plot of the clusters found in the Owens Valley region.

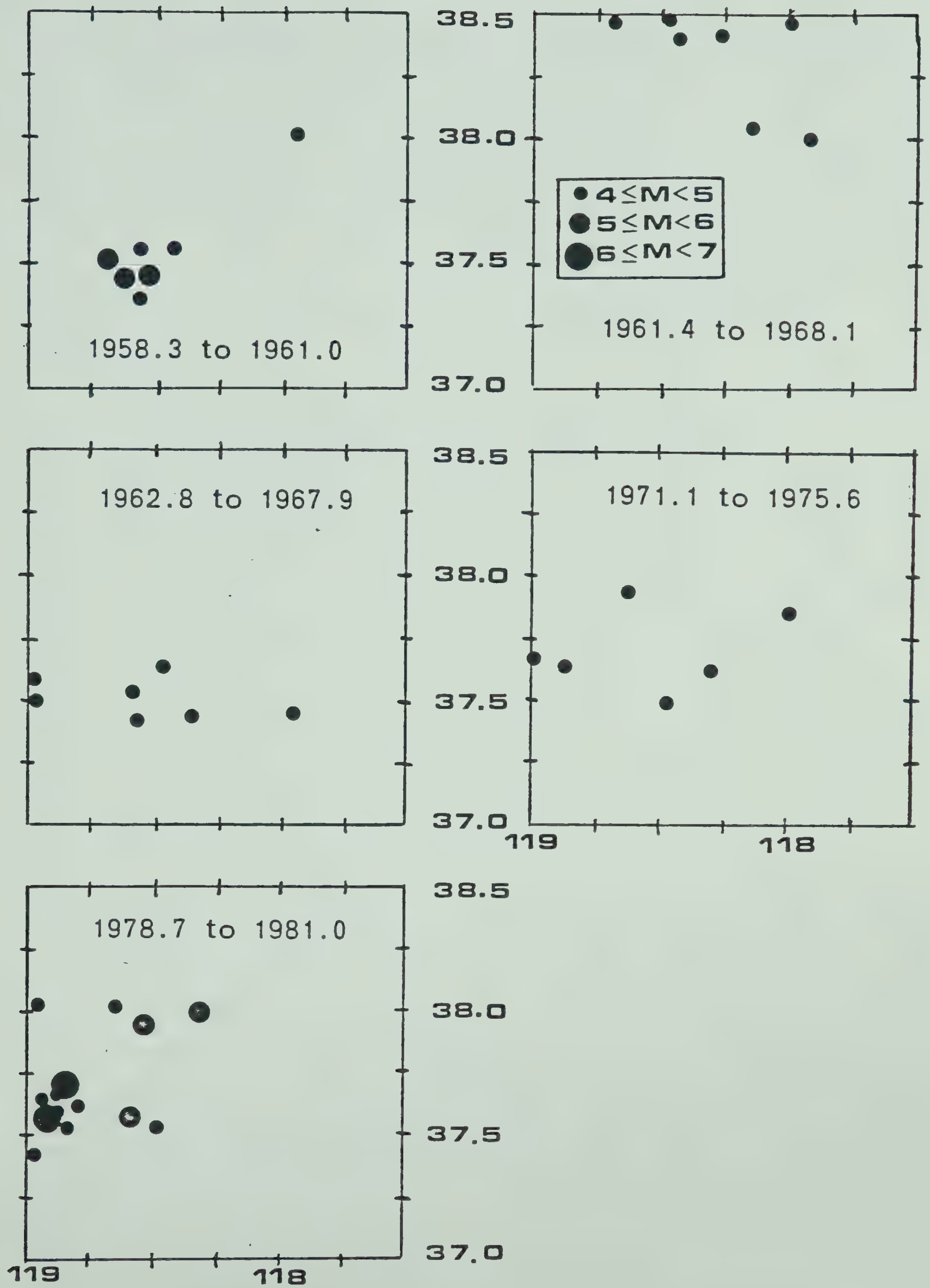


Figure 3.14b gives a continuation of Figure 3.14a.

shows a large spread in distance and time compared to the other clusters. The cluster which includes the Owens Valley event has just the opposite characteristics, most members, largest energy release, and it is one of the most compact clusters in time and space. There are no other similar quiescent → activation periods for the Owens Valley sequence.

3.7.6.2 Clusters in the Southern Sierra region

The Southern Sierra region (Figure 3.1) is an area of much lower seismicity rate than is Owens Valley. Figure 3.15 gives a latitude - longitude plot of clusters in the region. The clustering parameters are the same as they were for Owens Valley. The clusters found are very weak containing few members (5 or 6) and are not very localized in space. There is one exception. The cluster containing the 1946 magnitude 6.3 earthquake contains 13 members. Two earthquakes in the cluster occur after the 6.3 event, both in 1948. The cluster is much more concentrated in space than are other clusters in the region. The main activity in the cluster occurs on the lineament north of the epicenter of the magnitude 6.3 event (see Figure 3.1). It is this group of epicenters which generates the swarm found in Part 1 for this region (see Figure 3.8).

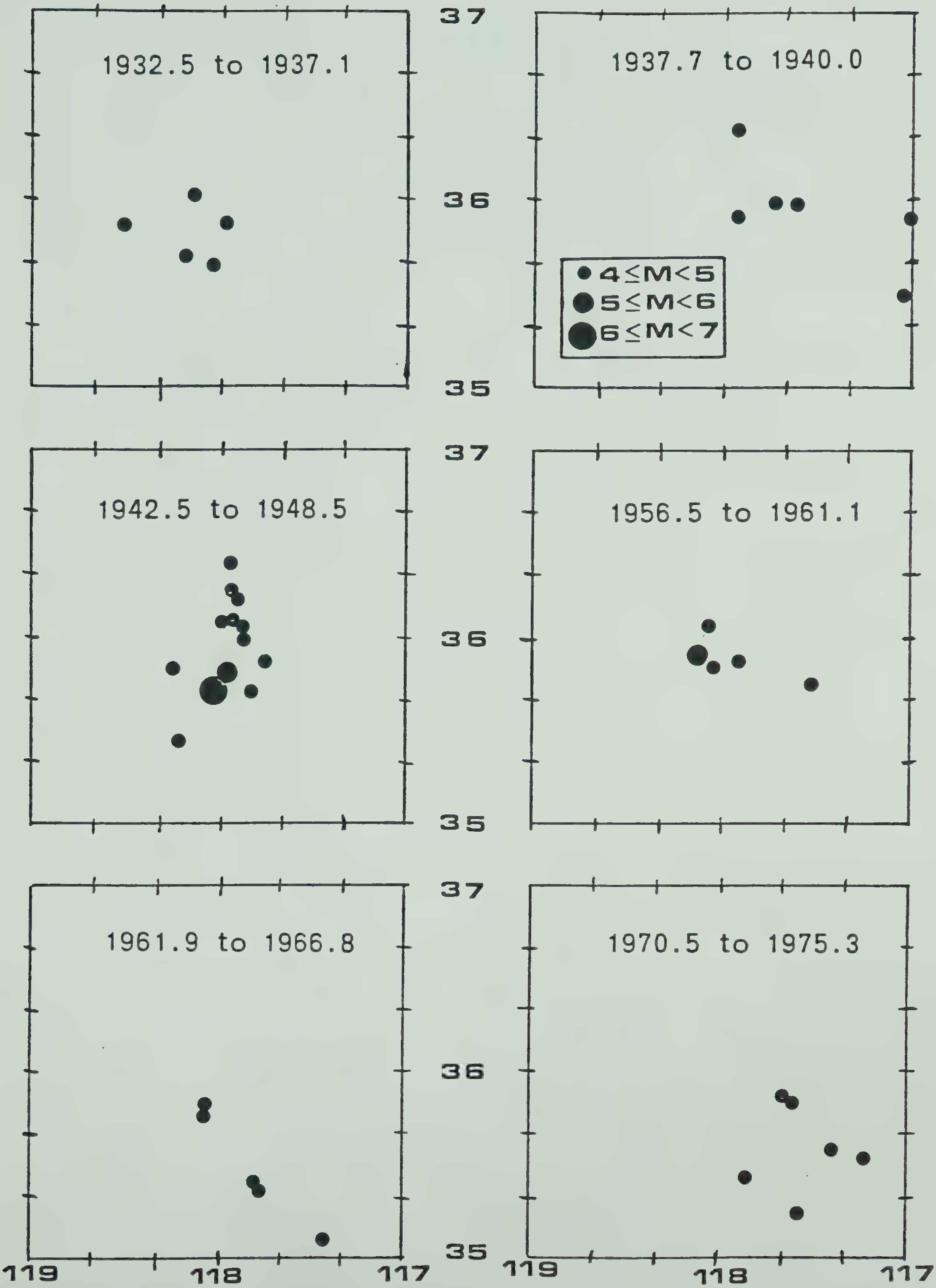


Figure 3.15 gives a latitude longitude plot of the clusters found in the the Southern Sierra region.

3.7.6.3 Clusters in the San Jacinto region

Figure 3.7 shows a clear quiescence followed by a swarm in the San Jacinto region during the last time windows of the catalogue. This pattern is of particular interest due to the suggested current high seismic potential of the region (for example Thatcher et al., 1975).

Figures 3.16a, 3.16b, and 3.16c gives a latitude - longitude plot of clusters for $M \geq 3.5$ in the region. A clustering normalization of 30 km. equal to 2 years was used. Figure 3.17 gives a time magnitude plot of the cluster members. The clusters display several interesting features:

1. The largest event in the region, the 1942 magnitude 6.5 event, was preceded by anomalous activity, although the suggested quiescence → activation pattern is not clear.
2. The quiescence → activation pattern occurring at the end of the catalogue is evident in figure 3.16.

Activity before 1937 is low and the clusters for those periods are not spatially concentrated. During the period 1941 to 1946 two very tight clusters were found. One of these clusters contains the 1942 magnitude 6.5 event. Fourteen of the 22 events in these clusters occur before the 1942 event and are concentrated on the same fault as the magnitude 6.5 event. The only other cluster which is so spatially concentrated is the last cluster.

The 1968 magnitude 6.4 event is not preceded by a similar pattern. The epicentral region of the 1968 event is quiet and no activation is evident before its occurrence.

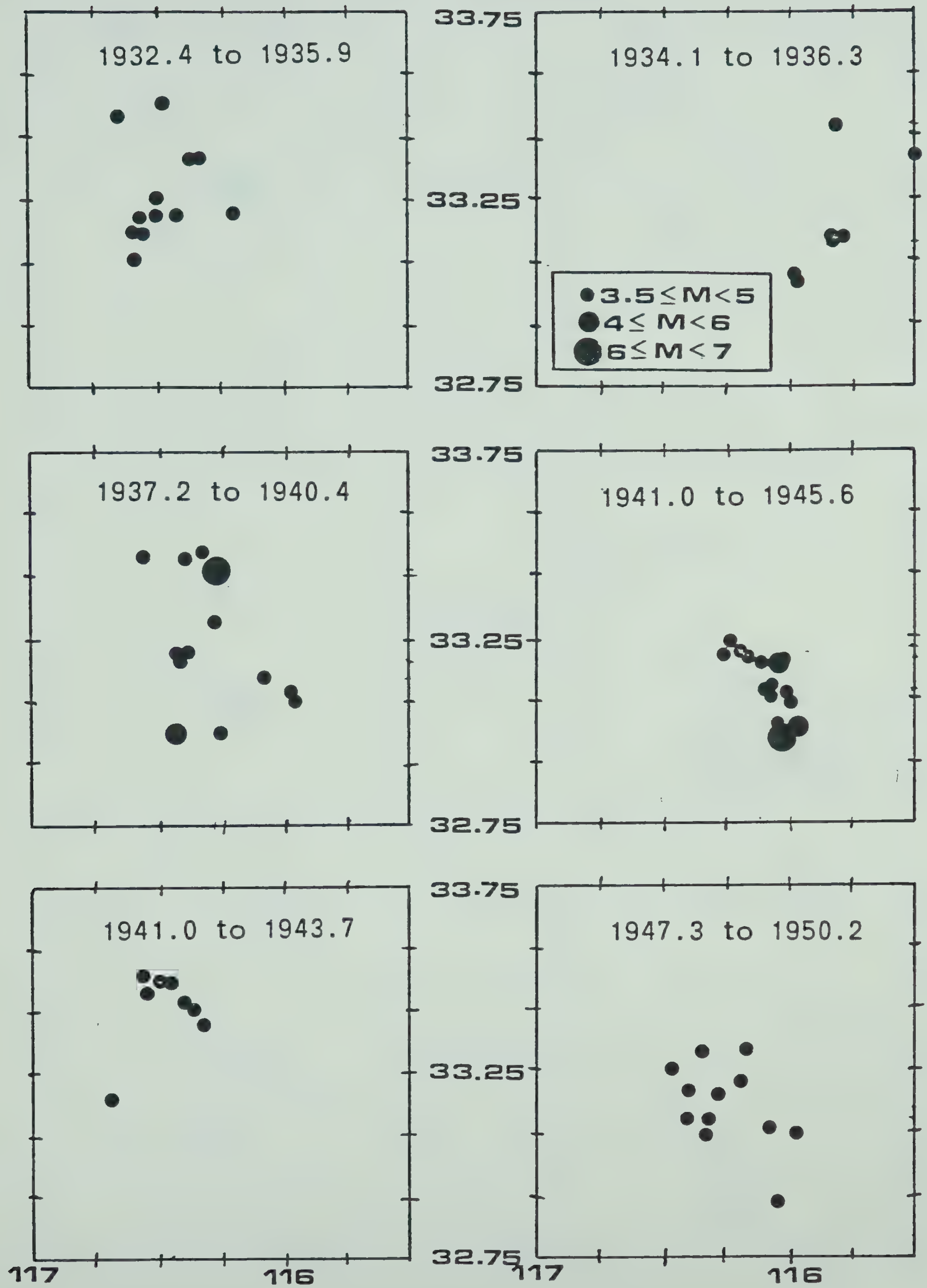


Figure 3.16a gives a latitude longitude plot of clusters found in the San Jacinto region.

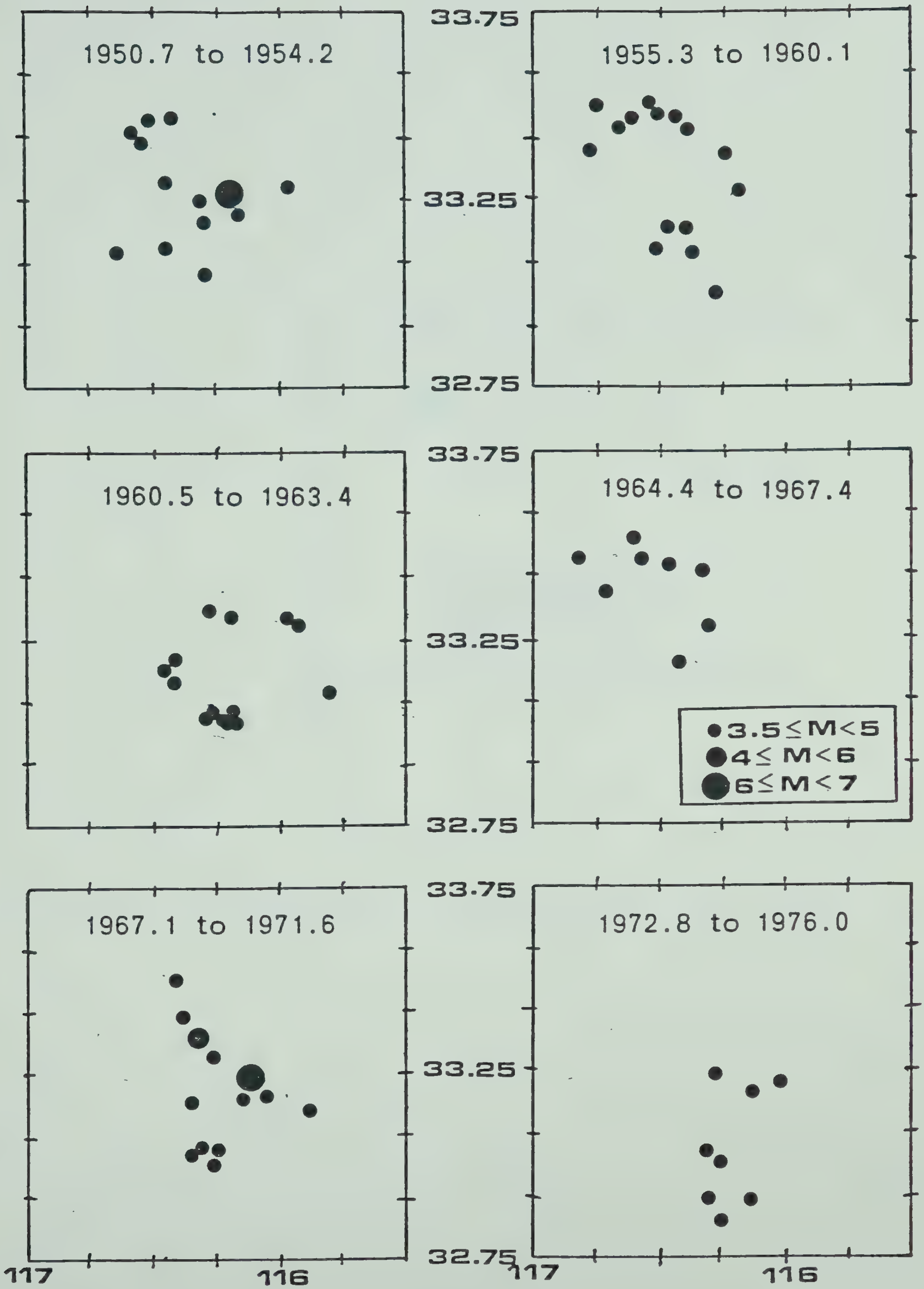


Figure 3.16b gives a continuation of 3.16a

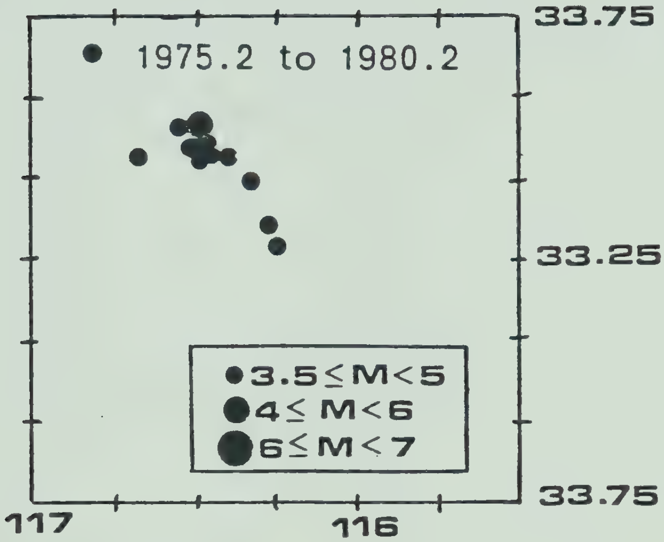
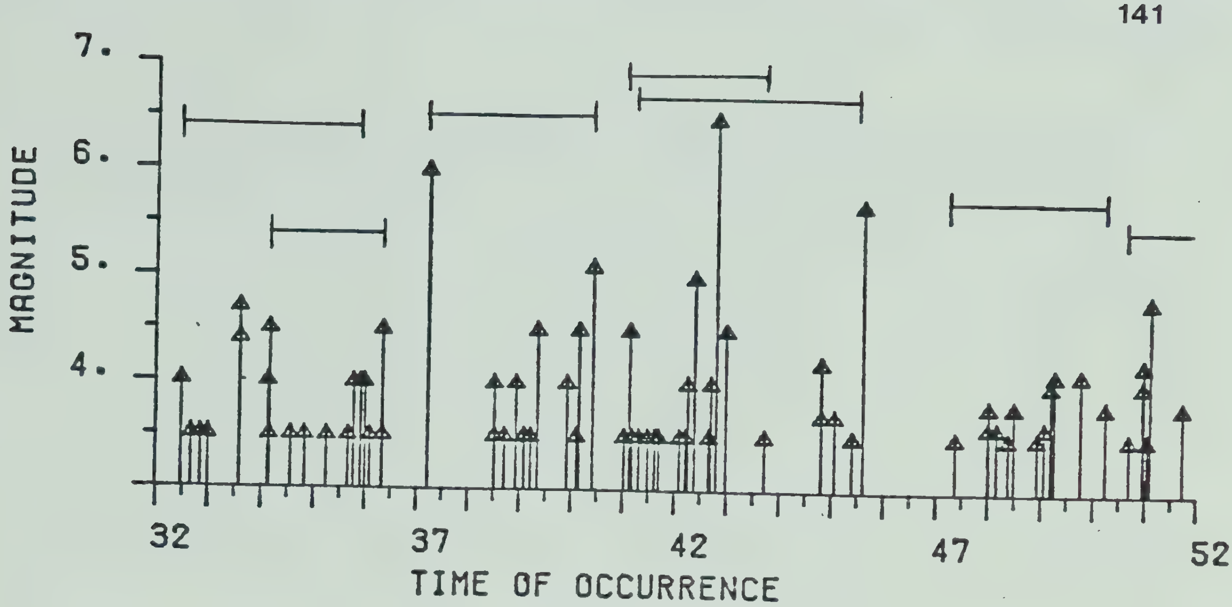
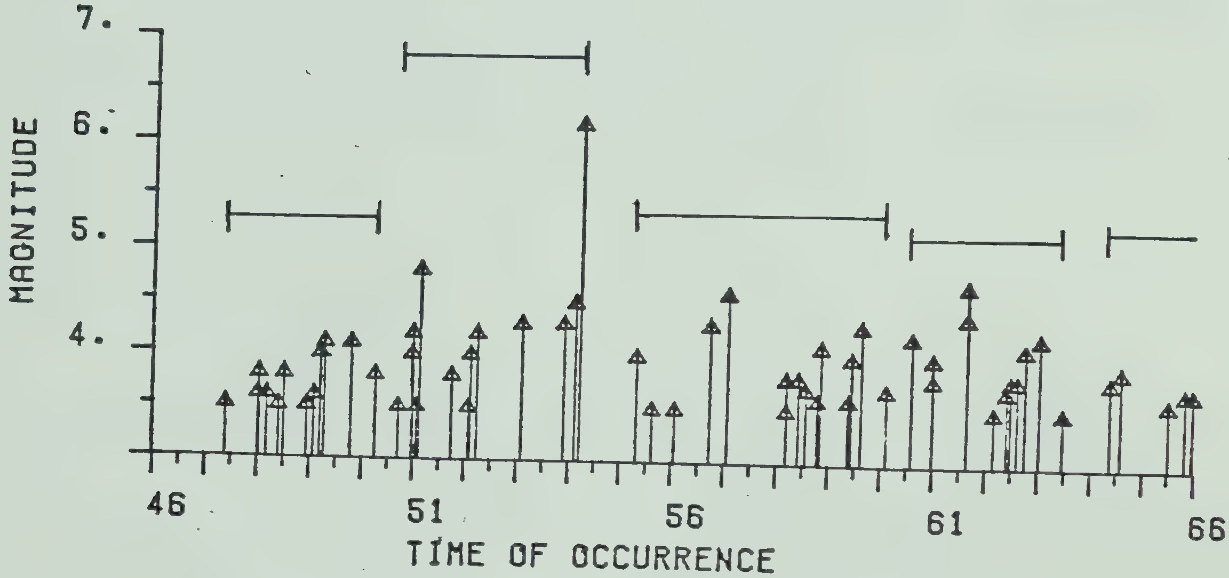
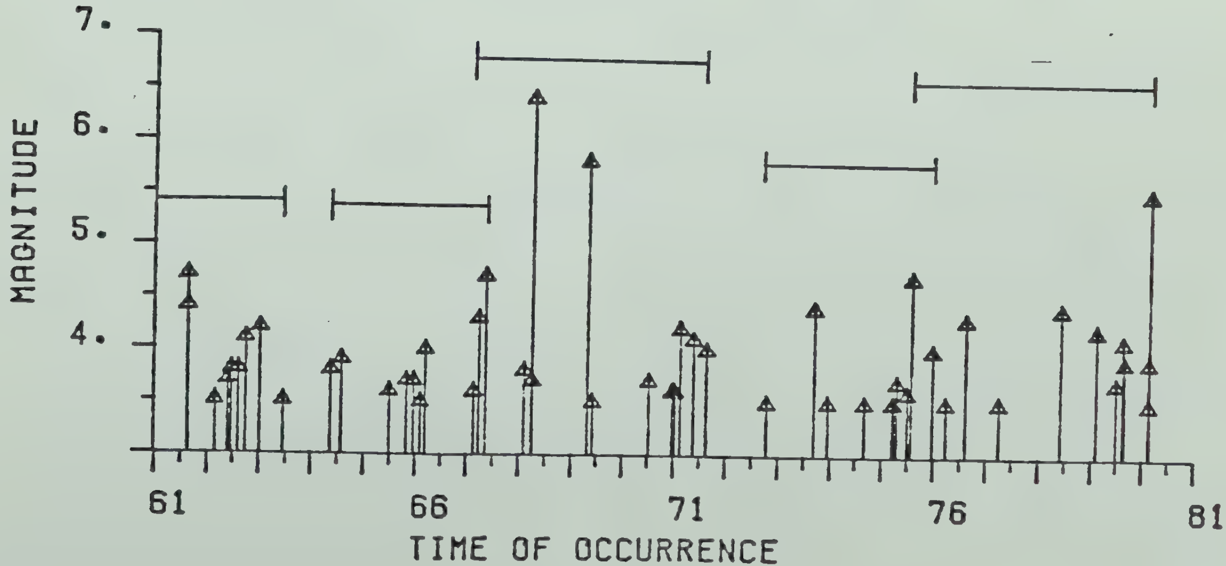


Figure 3.16c gives a continuation of 3.16a and 3.16b.

Figure 3.17 gives a time magnitude plot for members of the clusters in the San Jacinto region. The times spanned by the clusters are indicated by horizontal bars.


MAGNITUDE
7.
6.
5.
4.
32
37
42
47
52
TIME OF OCCURRENCE


MAGNITUDE
7.
6.
5.
4.
46
51
56
61
66
TIME OF OCCURRENCE


MAGNITUDE
7.
6.
5.
4.
61
66
71
76
81
TIME OF OCCURRENCE

Interestingly the only events in the clusters which occur near the future epicenter during the period 1954 to 1968 form a "doughnut pattern" around the epicenter. The doughnut pattern is formed by events occurring from 1961 to mid 1963, seven years prior to the major event. The entire fault system is very quiet after 1968 until 1975 when the spatially concentrated activity begins on the north section of fault. Events in the cluster occurring from 1972 to 1976 are mostly on the fault system to the southwest. The spatial concentration of events in the last cluster looks superficially very similar to the activity preceding the 1942 major event.

3.7.6.4 Other regions

In the preceding three examples the quiescence → activation patterns observed in part 1 are also visible in the clusters. This was true for other areas also. Clusters in the Imperial Valley region shows the quiescence → activation pattern before the 1979 magnitude 6.6 event. Similar results are seen in the Santa Barbara region before the magnitude 6.4 San Fernando event and in the Ensenada region before the 1956 magnitude 6.8 San Miguel event.

Regions in which the quiescent → activation pattern were not observed show similar cluster characteristics. Clusters in the Kern County and Desert Hot Springs regions show no obvious patterns before the magnitude 7.2 Kern County event and the magnitude 6.5 Desert Hot Springs event.

The low seismicity regions in the south, North Mojave, Riverside, Elsinore, and Los Angeles Basin have few cluster centers.

3.7.7 Density representation of clustering

Other means of characterizing the clustering of earthquakes are possible such as the pseudo density function $\hat{d}$ defined in section 3.6.4.3. Figures 3.18, 3.19, 3.20, and 3.21 show examples of $\hat{d}$ for mainshocks with $M \geq 3.5$. Here $\hat{d}$ is defined in the following way:

$$D_{ni} = \frac{1}{2}\sqrt{\frac{(x_n - x_i)^2}{x_0^2} + \frac{(y_n - y_i)^2}{y_0^2}} \qquad 19$$

where Dni is the distance from the nth event to its ith nearest neighbor and is given by

$$\hat{d}(X_k) = 10\left(13 / \sum_{i=1}^{13} D_{ki}\right) \qquad 20$$

x refers to latitude and y refers to longitude. The catalogue of mainshocks was divided into 6 year intervals and the function $\hat{d}$ was computed for these intervals. Specifically $\hat{d}$ is the density function of section 3.6.4.3, without time, calculated for 6 year time intervals with k=13. The distance Din is normalized so that 100 km. equals one unit of distance in the equation.

Figure 3.18 gives an average of $\hat{d}$ for Southern California using sliding 6 year time windows stepped one year. The average was constructed as follows: $\hat{d}$ was calculated for the six year intervals from 1932 to 1981. The

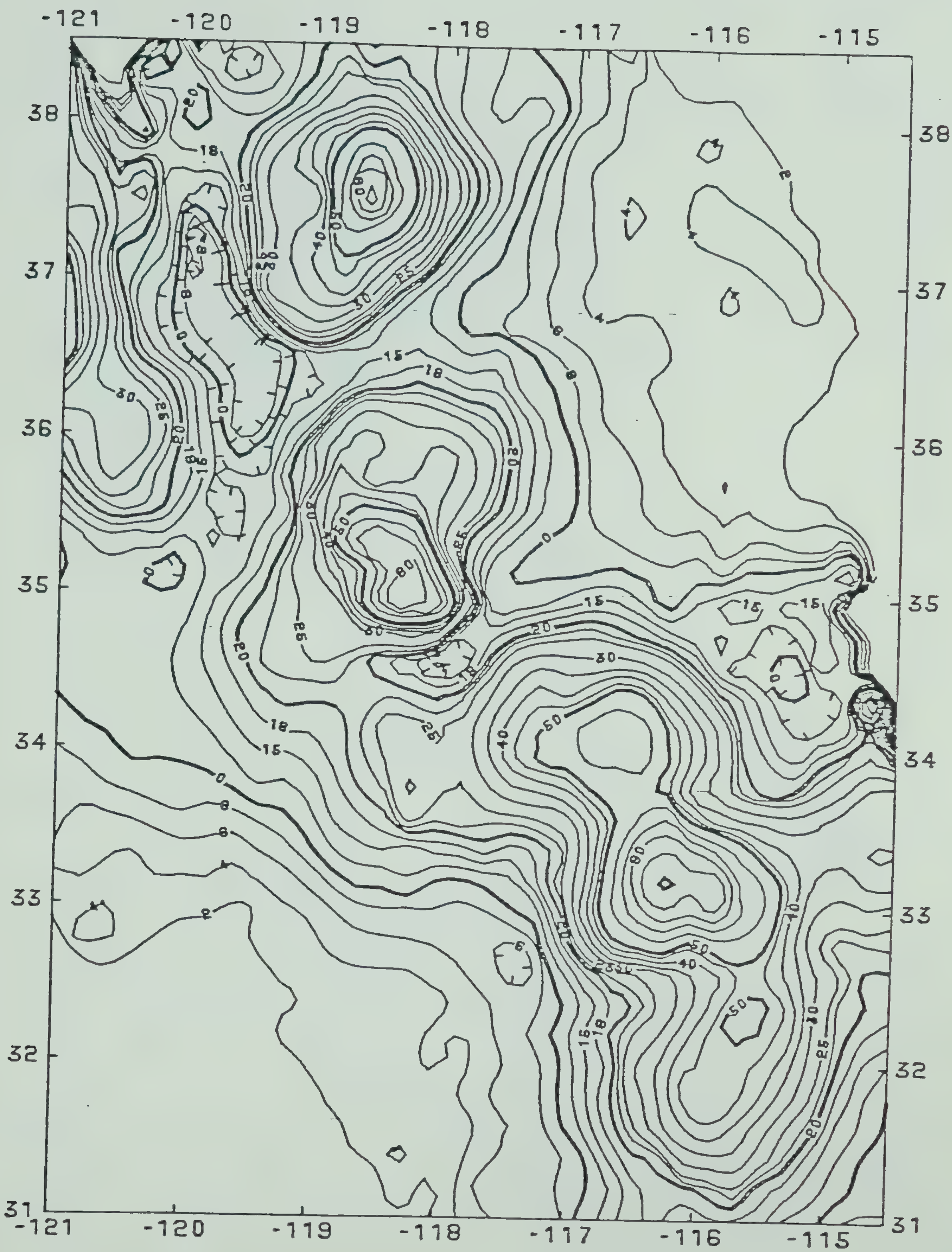


Figure 3.18 gives a contour plot of the average $\tilde{\sigma}$ for the Southern California region using sliding 6 year intervals stepped one year. Anomalous values of $\tilde{\sigma}$ have been eliminated from the average.

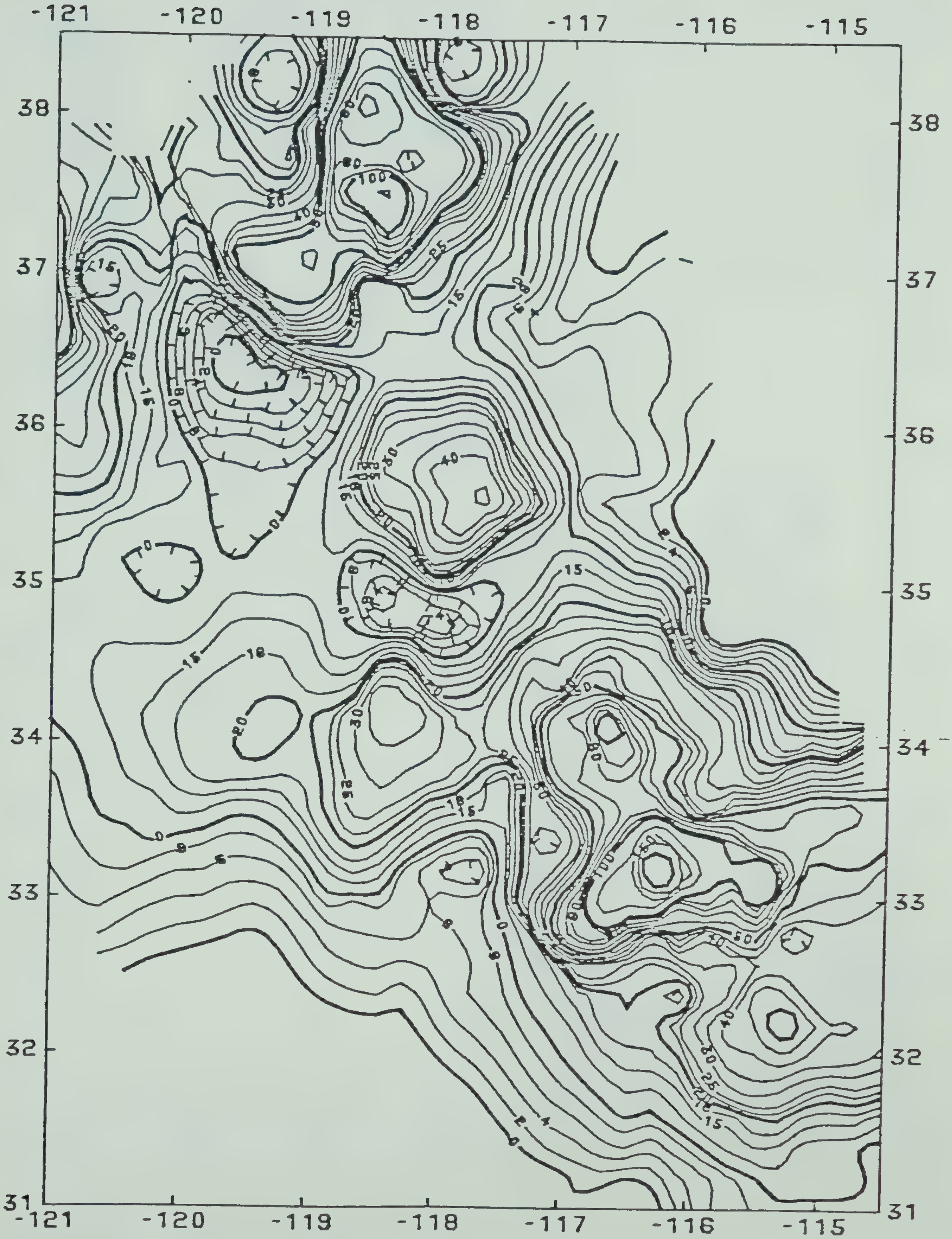


Figure 3.19 gives a map of $\tilde{\sigma}$ for the years 1945 to 1951.

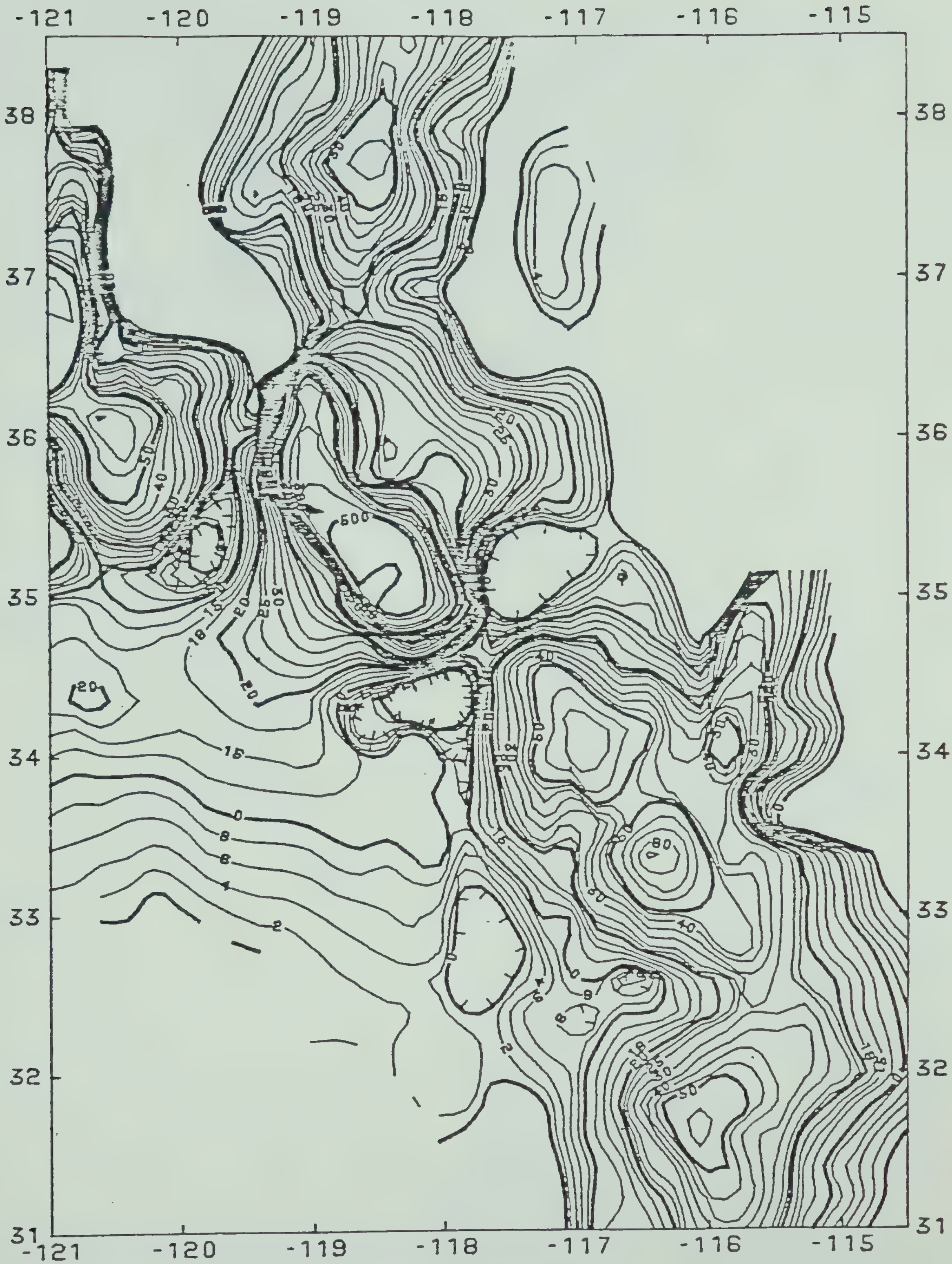


Figure 3.20 gives a map of $\tilde{\sigma}$ for the years 1950 to 1956.

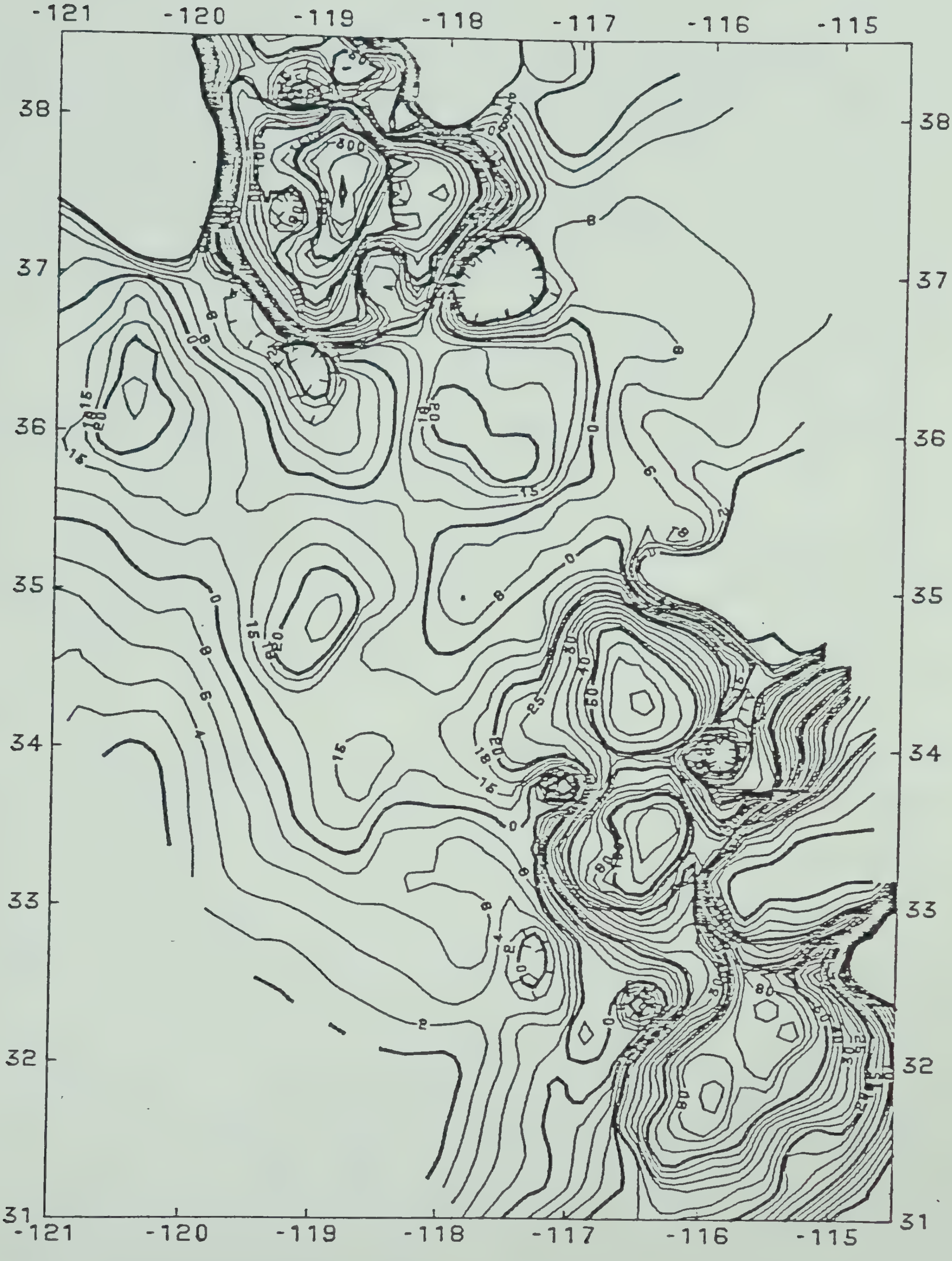


Figure 3.21 gives a map of $\hat{\sigma}$ for the years 1975 to 1981.

values of $\hat{a}$ at the epicentral locations were used to estimate values of $\hat{a}$ at regular grid points spaced .25°. These estimates were bases on weighted averages of the epicentral points using a weighting factor which decreased as 1/(distance from the grid point)^4. An overall average for each grid point was constructed. The average values were then recomputed after eliminating all values found to be anomalous. Anomalous values were defined as values at the grid points which differed by two standard deviations from the grid point means. The Surface II graphics package, as implemented on the University of Alberta computer, was used to generate the contour maps and grid estimates (Sampson, 1975).

Figure 3.19 and 3.20 give maps of the $\hat{a}$ for the periods 1945 to 1951 and 1950 to 1956. Changes in $\hat{a}$ in the vicinity of the magnitude 6.8 1956 San Miguel earthquake can be seen. During the period 1945 to 1951 (Figure 3.19) $\hat{a}$ in the vicinity of the future epicenter (31.5 latitude , -116.0 longitude) is low, approximately 10 compared to a long term average of 40, with no pronounced peak in the region. The period 1950 to 1956 (Figure 3.20) shows a distinct peak of 65 in the same region. The large peak in the center of the region in Figure 3.20 is due to activation following the 1952 Kern County earthquake. It should be noted that the catalogue is not complete for M≥3.5 for early years in the vicinity of the San Miguel earthquake. This makes comparison between the early values of $\hat{a}$ and the overall average

unreliable.

Figure 3.21 gives a map of $\hat{\sigma}$ for the last time period of the catalogue (1975 to 1981). The activations preceding and following the Owens Valley and Imperial Valley earthquakes are evident. Recent activation in the San Jacinto region, although not as pronounced, can also be seen. The central region of Southern California appears to be "quieter" than usual. Other representation using functionals similar to $\hat{\sigma}$ are possible.

3.7.8 Conclusions part 2

The results of cluster analysis support the hypothesis that mainshocks (as defined here) form a "quiescence → clustered" pattern before many of the largest earthquakes in Southern California. Although the number of cases presented is too few for definite conclusion, there is a suggestion that the activation stage of the quiescence → activation pattern is characterized by the strongest clusters in the region (ie, the most concentrated clusters in space and time). Some statistical model of regional clustering is needed to test this. A model that provides a quantitative definition of what constitutes anomalous activity (activation and quiescence) for a region is needed. Cluster analysis will be useful in developing such a model.

Although cluster analysis is computationally expensive and the results are tedious to interpret, it appears to be a useful tool in the analysis of earthquake sequences. One of

the obvious advantages of cluster analysis over the analysis presented in Part 1 of this chapter is that, cluster analysis does not require an a priori regionalization as did the swarms pattern (i.e. the regionalization given in Figure 3.1).

3.8 Concluding remarks

If aftershocks and foreshocks are removed, using the simple definition of Table 2.1, the catalogue of residual events ("mainshocks") in Southern California with $M \geq 4.0$ still shows a significant degree of clustering on time scales of 3 to 6 years and distances of the order of 100 km. The possibility of clustering on larger time and distance scales could not be tested due to the short length of the catalogue. The hypothesis that earthquakes are interconnected over large time and distance scales is supported by the calculations presented here.

Anomalous groups of the mainshocks, swarms or concentrated clusters, preceded by quiescence were found to precede, most but not all, large events in the Southern California region. This quiescent → activation pattern in mainshocks is similar to some patterns observed for all earthquakes by other investigators. The time scale of the quiescence → activation pattern for mainshocks is much longer than the usual proposed patterns.

The separation of mainshocks from aftershocks as described here can be viewed as a strong filtering process which removes short period variations in the catalogue. Short term anomalies may obsure longer term variations. The removal of aftershocks, as defined here, from the catalogue made the identification of the quiescence → activation clearer. The distribution of events in the residual catalogue is closer to a Poisson distribution than is the distribution of events in the highly clustered catalogue which includes aftershocks. Deviations from Poisson behavior in the residual catalogue are, thus, more statistically significant than in the "complete" catalogue. This technique may be useful elsewhere.

To my knowledge, this work is the first attempt at applying the mathematical technique of cluster analysis to the study of earthquake sequences. The analysis presented here obviously leaves many gaps. It does, however, show the usefulness of the technique in the study of anomalous seismicity patterns.

Cluster analysis is particularly adept at picking out regional periods of anomalous activity and quiescence. It appears that the technique would be very useful in the location and study of seismic gaps in, for example, Middle America. Another possible use of cluster analysis would be in the development of a statistical model of the clustering of earthquakes in space and time. Such a model requires some physical basis.

Obvious improvements can be made in the cluster analysis techique presented here. More physical metrics which also incorporate the earthquake magnitudes could be constructed. Also, a more physical means of determining the normalization of time with respect to distance in needed.

4. Bursts of seismicity near the Cocos - North American Caribbean triple junction

4.1 Introduction

Three related seismicity patterns, precursory to the largest earthquakes in a region, have been proposed by Keilis-Borok et al. (1980a,b,c). These three patterns, collectively termed "bursts of seismicity", consist of the abnormal clustering of earthquakes in time, energy and space before a major earthquake and were briefly described in Chapter 1. This chapter will apply the patterns; pattern B, pattern Swarms, and pattern Sigma, to retrospective prediction of large earthquakes which have occurred in the vicinity of the Cocos - North American - Caribbean triple junction.

Pattern B, "bursts of aftershocks", consists of a medium magnitude mainshock which has an anomalous number of aftershocks concentrated at the beginning of the aftershock sequence. Pattern Swarms consists of a "swarm" of mainshocks where a "swarm" is defined as a group of moderate events concentrated in space and time and occurring during a time interval when the overall seismicity is not below average. Pattern Sigma consists roughly of an increase in the cummulative seismic energy released, to the 2/3 power, in a sliding time window. Sigma is roughly proportional to a summation of the fracture areas of the earthquake sources

(Keilis-Borok and Malinovskaya, 1964). Pattern Sigma is identified as a peak in this summation.

In algorithmic form the patterns have been applied with some success to many regions, a review of which, is given in Section 1.6.3. Bursts of seismicity patterns are regional in nature. As such they do not indicate the location of the future earthquake precisely. In most cases the patterns take place at some distance (hundreds of kilometers) from the mainshocks which they precede. Accordingly, the occurrence of the patterns may be interpreted as indicating a significant increase in the probability of strong earthquake somewhere in the region within the next several years (Keilis-Borok et al., 1980b).

There is no adequate physical theory to explain the occurrence of these patterns. Various speculations suggest that non-linear features of "friction" on fault zones (Barenblatt et al. 1981) may play a role. Other suggestions include the obvious one that an increased density of small asperities (Kanamori, 1981) appears on the fault zone. The large separation between the patterns and the subsequent large event may indicate that both patterns and large event are symptoms of the same, as yet not understood, underlying cause.

4.2 The data set

For the Middle America region the only earthquake catalogue available in machine readable form was the NOAA PDE catalogue covering the years 1898 through 1979. The PDE listings for January, 1980 through July, 1980 were added to this. The nature of the NOAA data set presents two non-trivial problems in homogeneity, completeness of recorded events in time and consistency of magnitude estimates in time.

4.2.1 Magnitude consistency

Since the NOAA catalogue is compiled from many data sources there is the possibility of inconsistencies of magnitude estimates. Different sources have used different magnitude scales. The following scales are present in the NOAA catalogue: Ml (the Richter 1935 magnitude scale), Ms (surface wave magnitude), Mb (body wave magnitude) and Mu (unspecified magnitude). The most commonly used magnitude for statistical studies is Ms. For this reason Ms is used in this study.

Unfortunately, the majority of events in the NOAA catalogue do not give an estimate of Ms. It is therefore necessary to attempt conversions of other magnitude estimates to the Ms scale. Bloom and Erdmann (1979) show that for statistical purposes Ml and Mu are roughly equivalent to Ms. After 1963 the most frequently recorded magnitude in the catalogue is Mb with no other magnitude

indicator given. It was necessary to make estimates of Ms from Mb for most earthquakes from 1963. This is complicated by saturation of the Mb scale relative to the Ms scale for larger magnitude events (Kanamori and Anderson, 1975, and Kanamori, 1977). Large errors in estimated Ms values are possible for individual events. This constitutes the single largest difficulty in the use of pattern Sigma which critically depends on accurate magnitude estimates.

Many authors have approached the problem of conversion of Mb to Ms (i.e. Geller, 1976; Bloom and Erdmann, 1979; Kagan and Knopoff, 1980). Here the approach taken by Kagan and Knopoff (1980b) is used. For the conversion of Mb to Ms an empirical curve of the average values of Ms for a given Mb was constructed using all earthquakes in the Middle America region which had an estimate of both Mb and Ms. The resulting relation is similar to those given by Kagan and Knopoff (1980) and Bloom and Erdmann (1979). It consists of two segments:

$$Ms = Mb \text{ for } Mb \leq 6.0$$

$$Ms = 1.5Mb - 3.0 \text{ for } Mb > 6.0.$$

For the analysis the following rules were followed for the selection of a magnitude estimate $\bar{M}$ for an individual event:

$\bar{M}$ = Ms, if Ms is present

If Ms not present:

$\bar{M}$ = Ml or Mu, if Ml is not present

If only Mb is present:

$\bar{M}$ is given by the above empirical relation.

4.2.2 Completeness in Time

The completeness of the catalogue limits the time window of the study. Studies of correlations between large events and smaller events above some magnitude M, are limited to the study of only those events which occur after the catalogue becomes relatively complete for events with magnitudes M and above. Table 4.1 gives the number of events in various magnitude ranges for 5 year intervals starting in 1900. The table includes events in the entire Middle America region. Before 1910 only events with reported magnitudes greater than 7.5 are present. Numbers of magnitude 6.5 to 7 events start to become significant after 1910, numbers of 5.5 to 6. events after 1930, and numbers of 5. to 5.5 events after 1955. The number of events with magnitudes less than 5 increases continually.

The patterns will be tested on the available data starting in 1930 with the understanding that the catalogue is not very complete for early years. The results for early years could be biased due to this.

4.3 The region

Figure 3.1 shows the region of interest along with the locations of large ($M \geq 6.0$) earthquakes listed for the area in the NOAA catalogue. The area is a region of complex

Year	Magnitude interval <4.5	5.0	5.5	6.0	6.5	7.0	7.0+
1900-1905	0	0	0	0	0	0	8
1905-1910	0	0	0	0	0	0	3
1910-1915	0	0	0	0	1	4	4
1915-1920	32	0	0	0	0	2	6
1920-1925	54	0	0	0	1	2	3
1925-1930	125	0	0	0	1	7	8
1930-1935	225	0	0	21	34	17	5
1935-1940	249	0	0	10	8	9	4
1940-1945	215	0	0	9	12	13	6
1945-1950	49	0	0	3	6	11	7
1950-1955	318	2	9	33	34	17	7
1955-1960	424	19	43	28	30	18	2
1960-1965	674	103	50	17	11	9	3
1965-1970	917	281	83	31	11	3	2
1970-1975	269	351	130	35	13	4	3
1975-1980	331	408	118	33	12	4	4

Table 4.1 gives the numbers of events listed in the NOAA catalogue in various magnitude ranges of $\bar{M}$ for five year intervals starting in 1900. The table includes events in the entire Middle America region.

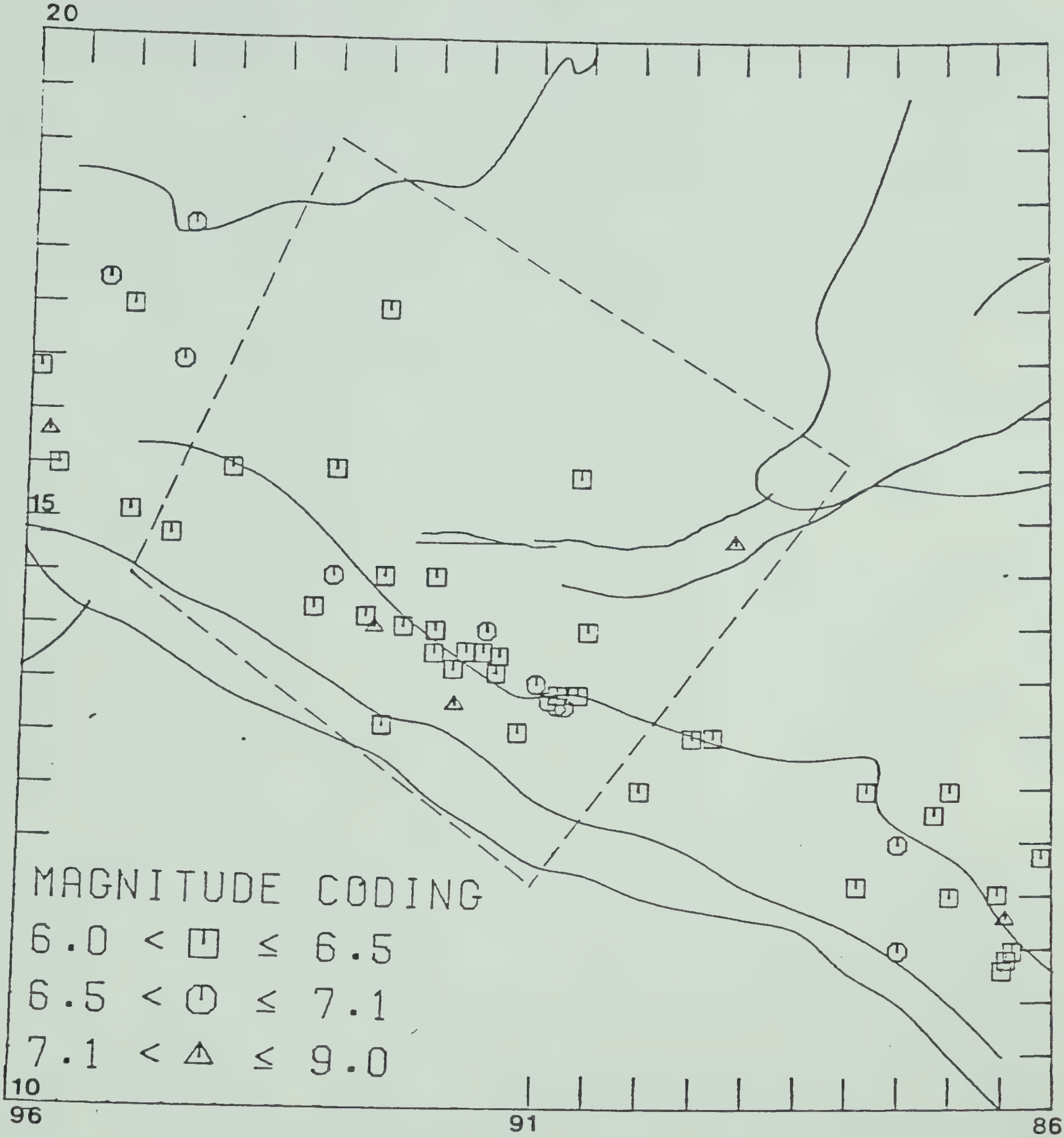


Figure 4.1 gives an outine of the area used in the bursts of seismicity study along with large earthquakes (M ≥ 6.0) listed in the NOAA catalogue for the area.

tectonics which contains segments of the boundaries between the Cocos and North American plates, the Cocos and Caribbean plates and the Caribbean and North American plates. The left lateral Motagua-Polochic fault zones define the Caribbean - North American plate boundary. The Cocos plate is subducting under the North American and Caribbean plates at a rate of 8.3 and 6.7 cm/year respectively at the Middle American trench near the triple junction (McNally and Minster, 1981). The nature and location of the Cocos - North American - Caribbean triple junction is poorly understood.

The plate boundary segments near the unstable triple junction have been the sites of numerous large shallow earthquakes. A list of large events ($M \geq 7.1$) which have occurred in the region since 1920 is given in Table 4.2.

The destructive February 4, 1976 Ms=7.5 Motagua earthquake occurred on the North American - Caribbean plate boundary. The Middle American trench region in the vicinity of the proposed triple junction (near the junction of the Middle America trench and the western extension of the trend of the Motagua - Polochic fault zones) has been the site of several $M \geq 7.0$ earthquakes since the turn of the century, the March 29, 1973 Ms=7.3 being the most recent. However, the segment of the trench northeast of the triple junction has not been the site of a historically recorded large ($Ms \geq 7.0$) earthquake. This, the southern end of the North American - Cocos plate boundary, is the site of the subduction of the Tehuantepec Ridge. This region may not have the potential

YEAR	MONTH	DAY	LAT.	LONG.	MAG.	DEPTH
1926	2	8	13.00	-89.00	7.1	0
1935	12	14	14.75	-92.50	7.3	0
1942	8	6	14.00	-91.00	8.3	0
1944	6	28	15.00	-92.50	7.0	0
1946	6	7	16.50	-94.00	7.1	100
1950	10	23	14.50	-91.50	7.1	0
1953	11	17	13.80	-91.80	7.4	0
1970	4	29	14.52	-92.60	7.3	33
1976	2	4	15.32	-89.10	7.5	5

Table 4.2 gives a list of large ($M \geq 7.1$) events occurring in the study region since 1920.

for large shallow earthquakes (Kelleher and McCann, 1976).

The segment of the Cocos - Caribbean plate boundary south-east of the triple junction has been the site of the largest earthquake in the region, the August 6, 1942 M=7.9 event. The segment of the trench just north-west of the site of the 1942 earthquake was the site of the Nov. 17, 1953 M=7.3 earthquake. Neither of these segments of the trench have ruptured since.

On the basis of historical seismicity several authors have identified the site of the 1942 earthquake and the segment of the trench south-east of the 1942 earthquake as a seismic gap of the first kind (Kelleher et al., 1973; McCann et al., 1979). The lack of large events in the region since 1942 has been clearly shown by Kelleher et al. 1973. McNally and Minster (1981) point out that estimated cumulative seismic slip in the same area has been low since the 1942 earthquake.

Earthquake recurrence intervals in the Middle America region have been found to be highly regular by several authors. Rikitake (1976) calculates the average return period for large shallow earthquakes in the Middle American region as 34.5 ±3.6 years. McNally and Minster (1981) estimate the average return period as 32.5 ±8.4 years on the basis of recorded data 1898-1979, and as 35.1 ±24.0 years on the basis of historical data 1542-1979 in the Oaxaca region. On the basis of estimated average return times and the low amount of cumulative seismic slip McNally and Minster (1981)

suggest that the Guatemala coast deserves special attention as a zone of presently high seismic risk.

Figure 4.1 shows that the region of highest seismicity in the area (at least for $M \geq 6.0$) is in the vicinity of the proposed triple junction. It is suggested here that the nature of the seismicity this complex region is indicative of the state of stress on the adjacent plate segments and that periods of high energy release and earthquake clustering in the area of the triple junction occur when one or more of the boundaries is near major failure.

4.4 Bursts of seismicity

It is first necessary to define some terms. Strong events are the set of largest events in a region, those with magnitudes above some threshold M_0. Earthquakes with magnitudes less than or equal to M_0 but greater than some minimum threshold $\bar{m}$ are termed the background seismicity. Relatively large shocks in the background seismicity, those with magnitudes between M_1 (a lower limit) and M_2 (an upper limit) are called main shocks. Aftershocks of major events are excluded from this classification.

Some rules are needed to analyze the success of a given pattern. If a pattern is valid then the occurrences of the pattern should show a marked tendency to fall within some time window preceding major events. The length of this time window may depend on several parameters such as the

magnitude of the strong events and the region under consideration. The length of this time window will be termed t_0. A success is then defined as the occurrence of a pattern or a sequence of patterns which is followed by a major event within time t_0. A failure will be defined as a major event which was not preceded by a pattern within time t_0. A false alarm is defined as an occurrence of a pattern which is not followed by a major event within time t_0.

4.4.1 Swarms

A swarm is defined as a group of earthquakes concentrated in time and in space and occurring at at time when the overall seismicity in the region in not below average. Some additional definitions (Keilis-Borok et al., 1980b) are needed.

--An aftershock is defined by the following: Consider two earthquakes with time sequence numbers i and j, i>j. The second earthquake is an aftershock of the first if the following conditions are satisfied: The distance between their epicenters is less than R(Mi); the time difference $ti-tj \leq T(Mi)$; and $Mj \leq Mi$, where T(M) and R(M) are empirical functions. These thresholds must be set a priori.

R(M)is defined as: R(M) = 75 Km.; T(M) is taken from Kellis-Borok et al.(1980.a): T(M) = 0.5 yr for $5.0 \leq M \leq 5.4$; T(M) = 1 yr for $5.5 \leq M \leq 6.4$ and T(M) = 2 yr for $M \geq 6.5$.

--N(t) = the total number of of earthquakes with $\bar{M} \geq \bar{m}$

occurring in the entire region within the time interval from t-s to t.

--$\bar{N}(t)$ = a running average of $N(t)$ from the beginning of the catalogue to time t.

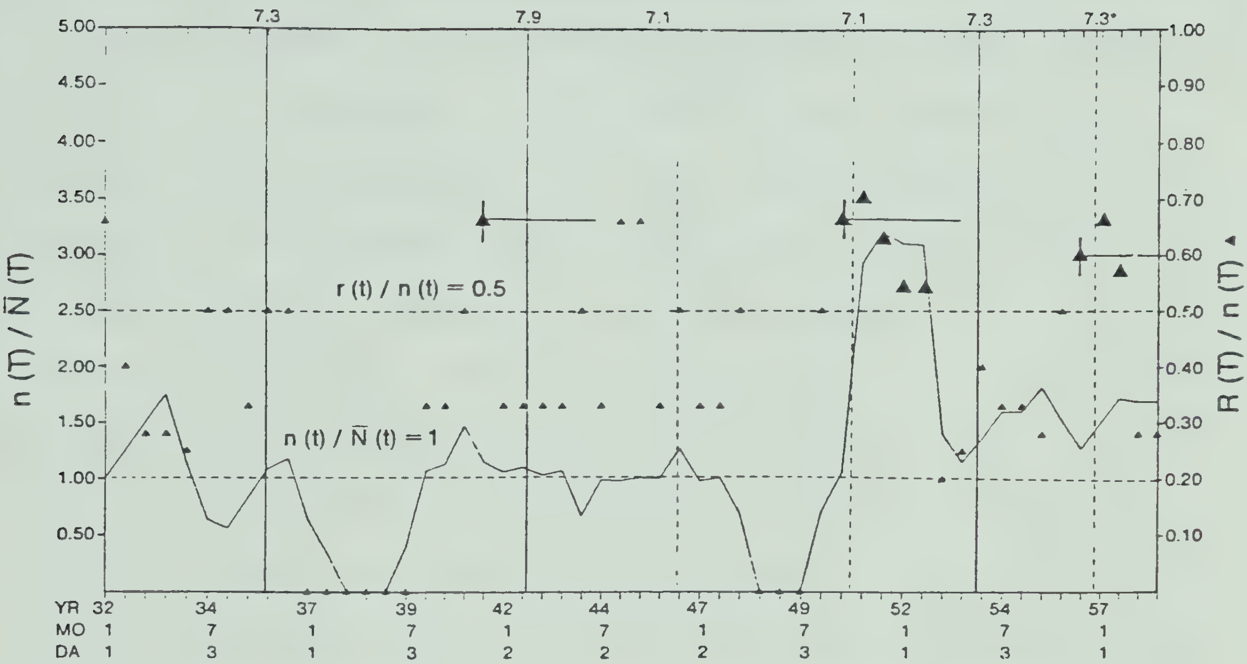
--$n(t)$ is the same as $N(t)$ but excludes aftershocks of strong earthquakes.

--$r(t)$ = the maximum number of earthquakes with $\bar{M} \geq \bar{m}$ occurring within the time interval extending from $t - s$ to t and occurring within an arbitrarily located rectangle R of dimension Δx by Δy in latitude and longitude within the region (i.e. $r(t)$ is a measure of the clustering of the mainshocks). The parameters s, Δx, and Δy are free parameters of the algorithm.

A group of earthquakes within a rectangle Δx by Δy is called a swarm if $n(t) \geq \alpha\bar{N}(t)$ and $r(t) \geq \beta n(t)$.

The following choices were made for the free parameters: $M_0 = 7.3$; $s = 2$ years; $\alpha = 1.0$; $\beta = .5$; $\Delta x = \Delta y = 1°$; and $t_0 = 3$ years. The values used for s, Δx and Δy are greater than those used in California (Keilis-Borok et al., 1980b). These values were increased due to the larger maximum magnitude of earthquakes and poorer epicentral locations of events in Middle America relative to those in California.

Since the minimum magnitude threshold of reported events in the catalogue changes significantly between 1950 and 1963 two separate time segments have been considered. The results are given in Figure 4.2. A swarm is said to have



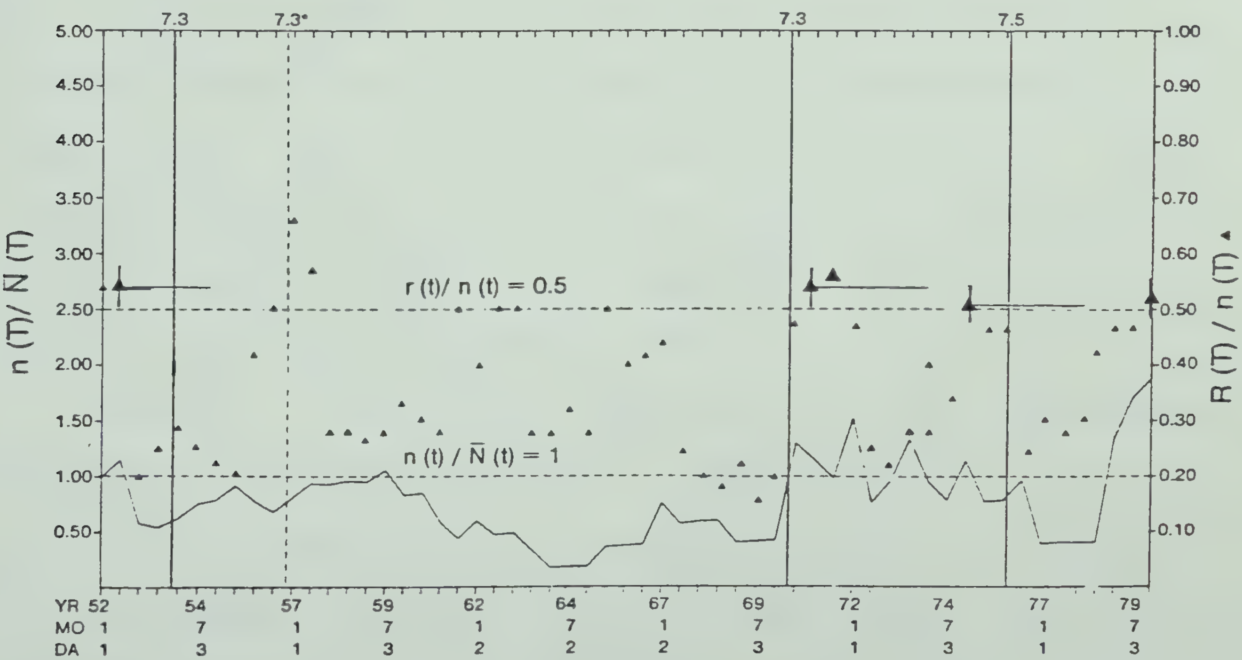


Figure 4.2 shows plots of $n(t)/\bar{N}(t)$ and $r(t)/n(t)$ for the periods 1932 to 1959 (a) and 1952 to 1980 (b). Large earthquakes ($M \geq 7.2$) are indicated by vertical lines. Dashed lines indicate slightly smaller events of events occurring just outside the study region.

occurred when both $n(t)/\bar{N}(t) \geq 1.0$ and $r(t)/n(t) > .5$. The occurrences of swarms are marked in the figures. For the period of time 1932 to 1950 there are not many events in the catalogue. However, using the available data several swarms are observed. The Dec. 14, 1935 M=7.3 event occurred too early to test the pattern. The first clear swarm occurred during the time window ending on January 1, 1941. This precedes the M=7.9 1942 event by approximately 1.5 years. After the 1943 large event the average number of events in the catalogue dropped and there were no swarms until the strong swarm during the time interval ending on January 1, 1951. The swarm continued during the next 2 years. The time between the beginning of the swarm and the occurrence of the M=7.3 event in 1953 is just less than 3 years. Activity either remained high or the character of the catalogue changes after the 1953 event. Figure 4.2a indicates a false alarm starting in mid 1955. This may be due to activation before the Oct. 24, 1956 M=7.3 event which occurred just south of the region considered here.

Figure 4.2b begins with the end of the swarm preceding the 1953 event. The false alarm in 1955 disappears as $n(t)/\bar{N}(t)$ is less than 1. There are no swarms from 1955 to 1970. The pattern fails to detect a swarm before the April 29, 1970 M=7.3 event. Activation due to this event produces a false alarm (it should be noted that much of this activation is due to a foreshock sequence that preceded the strong event by a day; if the preceding time window had

included this foreshock activity a swarm would have been detected). The February 4, 1976 M=7.5 Motagua earthquake is preceded by a swarm occurring during the time period ending January 1, 1975. Interestingly, a swarm occurs at the end of the data set in the time window ending January 1, 1980. All swarms occur within a degree of the intersection of the trend of the Motagua-Polochic fault zones and the Middle America Trench.

4.4.2 Bursts of aftershocks

According to Keilis-Borok et al. (1980a, 1980b) the occurrence of a "burst of aftershocks" (pattern B) indicates a significant increase in the probability of a strong ($M \geq M_0$) earthquake in the same region within the next t_0 years. "Pattern B" consists of a mainshock in the medium magnitude range with an anomalous number of aftershocks concentrated at the beginning of the aftershock sequence. Specifically the pattern is a mainshock with magnitude Mi such that $M_2 \leq Mi \leq M_1$ and $bi(e) \geq \bar{B}$, where bi(e) is the number of aftershocks in the first e days following the mainshock and $\bar{B}$ is a threshold above which bi(e) is considered to be anomalous.

The number of aftershocks was counted with e=2 days. Table 4.3 gives a listing of all mainshocks with 4 or more aftershocks in the 2 day period following the mainshock. Also listed are the large earthquakes ($M \geq 7.2$) occurring in the region after 1950. Unfortunately, the NOAA catalogue

Anomalous numbers of aftershocks and
large earthquakes

year	month	day	M	lat.	long.	b(2days)
1950	10	23	7.1	14.50	-91.50	14
1953	11	17	7.4	13.80	-91.80	0
1970	4	29	6.3	14.62	-92.69	6
1970	4	29	7.3	14.52	-92.60	19
1973	6	7	6.2	14.28	-92.01	6
1976	2	4	7.5	15.32	-89.10	2
1979	10	27	6.8	13.83	-90.88	4

Table 4.3 gives a listing of all mainshocks ($M \geq 6.0$) with 4 or more aftershocks in the 2 day period following the mainshock. Also listed are the large ($M \geq 7.2$) earthquakes which have occurred in the region since 1950.

is not sufficiently complete to identify anomalous levels of aftershock activity with any reliability. No estimate of magnitude is provided for most aftershocks contained in the catalogue. The listing in Table 4.3 includes all aftershocks: no lower magnitude threshold was used.

Even with the small numbers of aftershocks the pattern scores well. If t_0 = 3 years, as was the case for the Southern California study, $\bar{B}$ = 6 aftershocks, and M_0 = 7.3 the pattern scores 2 successes, one false alarm, and one failure to predict. The false alarm is generated by the magnitude 7.1 event which occurred on Oct. 23, 1950, three years and 24 days before the magnitude 7.3 event in 1953. If t_0 is relaxed slightly to include an extra month the pattern would score 3 successes, no failures and no false alarms. If $\bar{B}$ is lowered to 4 aftershocks one more alarm occurs at the end of the catalogue.

4.4.3 Sigma

Pattern Sigma consists of a peak of the function

$$\mathcal{S}=\sum_{t-s}^{t} C\,10^{D(M-E)}$$

where the summation is taken over a sliding time window of duration t-s to t and over the magnitude range from M_1 to M_2. This pattern was first introduced by Keilis-Borok and Malinovskaya (1964). Specifically pattern Sigma is said to have occurred if $\mathcal{S}(t) \geq \bar{\mathcal{S}}$, where $\bar{\mathcal{S}}$ is some threshold value

which defines anomalous activity. Since $\S$ is an exponential function of magnitude its successful application depends on consistent and accurate relative magnitude estimates over the entire length of time used in the study. For the NOAA catalogue this condition, if met at all, probably only occurs after 1963. Regardless of this, the pattern is tested here on the data from 1932.

The values of the parameters were selected as above: s=2 years; M_0=7.3; M_1=5.0; M_2=7.0; C=1.0, D=.91 and E=4.5. Figure 4.3a and 4.3b give plots of $\S$ from 1932 to 1960 and form 1960 to July 1980. The success of pattern $\S$ depends critically on the choice of $\bar{\S}$.

One feature apparent from a study of $\S$ is that values of $\S$ before 1963 (the time of significant improvement in the catalogue) are, in general, much larger than values after 1963 even though the catalogue is more complete after 1963. This could be due to higher seismic energy release before 1963 or to over-estimates of earthquake magnitudes before 1963 relative to those after 1963. The latter could result if Ms was systematically under-estimated by the empirical formula of Section 4.2.1. Even with the low quality data, pattern Sigma works well for early years. The 1936 event occurs too early to test the pattern, there is a peak before the 1943 event, a possible false alarm in 1946 which can be eliminated with an appropriate choice of $\bar{\S}$, and the largest peak precedes the 1953 event. After the 1953 event seismic energy release remains relatively high, producing one or two

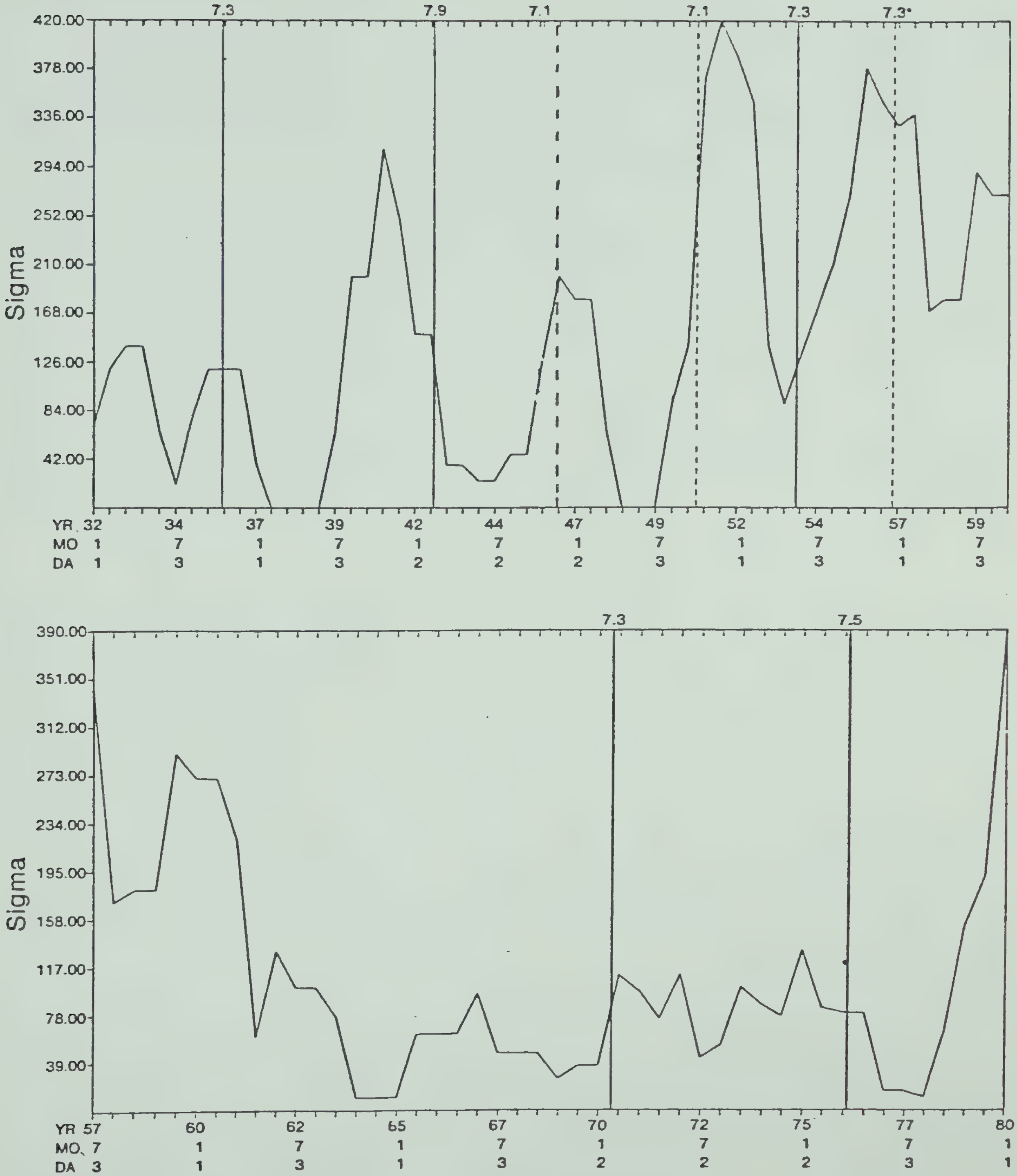


Figure 4.3 shows plots of S for the periods 1932 to 1960 (a) and 1957 to 1980 (b). Large events are indicated as in Figure 4.2.

false alarms between 1954 and 1960 depending on the choice of $\bar{\S}$.

The results after 1963 are complicated by the possible inconsistencies between early and later magnitude estimates. If the luxury of choosing separate $\S$ before and after 1963 is allowed, the the following conclusions can be drawn: The 1970 event is not preceded by a peak (again activation indicated in the time windows following the event are in part due to the foreshock sequence mentioned above) and a weak peak precedes the 1976 event. The catalogue ends with a strong peak (even by pre-1963 standards).

4.5 Concluding remarks

Given the quality of the data set, the patterns seem to score well. Without including the 1936 event pattern Swarms scores 3 successes (1943, 1953, 1976) one failure (1970) and one possible false alarm (1956, excluding the false alarm generated by the foreshocks and aftershocks of the 1970 event). With the most optimistic choice of $\S$ on two separate time intervals pattern Sigma scores the same successes and failures as does pattern swarms with a longer false alarm time in 1956. In the most optimistic case pattern B scores 3 successes, no failures and no false alarms. All occurrences of the patterns are close to the proposed location of the unstable triple junction. Success of the patterns may be due to the complex nature of the tectonics in the triple

junction area.

The last available data indicate the recent occurrence of both pattern Swarms, pattern Sigma, and possibly pattern B in the region due to events occurring during 1979 and early 1980. Based on the past history of the region and the lack of large events on some segments of the trench for more than 30 years, this region warrants attention as a possible site for the occurrence of a large ($M \geq 7.3$) shallow earthquake within the next 3 years.

It should be noted, however, that this suggestion is based on data which is not of as high a quality as that used for similar studies in California. The NOAA PDE file is incomplete for low magnitudes in its early years and magnitude estimates are, to say the least, inconsistent. Even the identification of the largest earthquakes in the region for the early years is subject to question.

For example the 1953 earthquake which was identified as having a magnitude of 7.3 in NOAA is listed as magnitude 7.1 by McNally and Minster (1981). This would then not qualify as a strong earthquake by the criteria used here. If the threshold for the definition of a large event is lowered to M_0=7.1 two other earthquakes would be added to the list of major events. Inclusion of these two events would improve the score for Sigma but not for swarms and bursts of aftershocks.

If the patterns are considered only for the relatively reliable data after 1963 the score for both Sigma and swarms

is one success, one failure and the current alarm while pattern B scores 2 successes. The comments presented here should be considered in the light of this and the fact that the patterns depend heavily on the inconsistent magnitude estimates.

5. Concluding Remarks

5.1 Summary

In this thesis several aspects of the clustering of earthquakes related to precursory seismicity patterns have been explored. Although there doesn't exist an unambiguous definition for aftershocks, the most obvious indications of earthquake clustering are mainshock - aftershock sequences. Using a simple definition for an aftershock, Chapter 2 proposes a self-similarity in this clustering, a scaling law for the occurrence of aftershocks. The scaling law suggests that number of aftershocks does not depend on the magnitude of the mainshock if the aftershocks are counted in a fixed magnitude range below that of the mainshock. This scaling law reduces the number of observations necessary for statistical studies of aftershock sequences and imposes a constraint on acceptable physical models of aftershocks. It is also suggested that foreshocks, on the average, generate larger numbers of aftershocks than do mainshocks. The scaling law supports the hypothesis that the distribution of irregularities in material strengths on a fault zone shows no characteristic scale.

In Chapter 3 the clustering of earthquakes on longer time and larger distance scales is explored. One way to look for longer term clustering is to first filter out the short term variations which could possibly obscure the picture.

This was the approach taken here. To study the possibility of long-term clustering "foreshocks" and "aftershocks" were removed from the catalogue using a simple boxcar filter definition for foreshocks and aftershocks. The purpose of the box car definitions was to remove the obvious short term clustering which sometimes precedes and followings a shock. In this study "mainshock" then refers to the largest event in one of these primary, box car defined clusters. "Aftershocks" and "foreshocks" are those events which occur respectively after and before the "mainshock" in the box car window. In this definition a "foreshock" - "mainshock" - "aftershock" sequence can be thought of as first order clustering which is perhaps imbedded in a hierarchy of higher order, longer term clustering.

It was demonstrated that even if foreshocks and aftershocks are elimiminated, through such a strong filtering process, the catalogue of residual events ("mainshocks") still shows weak but significant clustering characteristics. The clustering is evident on time scales of the order of 3 to 6 years and distance scales of the order of 100 kilometers for mainshocks with magnitudes greater than or equal to 4. The possibility of longer term clustering could not be studied due to the short length of the catalogue.

The hypothesis that there are related events in the catalogue over long periods of time and large distances, such as is suggested by the stochastic branching models of

earthquake occurrence, is not in conflict with the observations presented here. If, for example, the Kagan and Knopoff branching model, in which all shocks can be thought of as "children" of those shocks occurring before them in time, is correct no length of box car would be sufficient to eliminate all dependent events. The Kagan and Knopoff model was discussed in Section 1.11. In such a model clustering would always be observed at time scales longer than the length of the boxcar. Unfortunately, the short length of available catalogues procludes the possibility of testing for the presence of such long term clustering.

Anomalous periods of clustering and anti-clustering in the sequence of residual events were juxtaposed with the occurrences of large earthquakes in the same region. Anomalous groups of residual events, swarms or concentrated clusters of "mainshocks", were found to precede, most but not all, large earthquakes in the Southern California region. The quiescence → activation pattern observed here for mainshocks is similar to precursory patterns observed by other investigators, although the time scale is in general longer.

Short term anomalies in the catalogue caused by aftershock sequences may obsure longer term variations. The removal of aftershocks, as defined here, from the catalogue made the identification of the quiescence → activation clearer. The distribution of events in the residual catalogue is closer to a Poisson distribution than is the

distribution of events in the highly clustered catalogue which includes aftershocks. Deviations from Poisson behavior in the residual catalogue are, thus, more statistically significant than in the "complete" catalogue. The technique used here may be useful elsewhere.

Chapter 3 also demonstrates that the mathematical technique of cluster analysis is useful in the study of anomalous seismicity patterns. It is suggested that the activation stage of the quiescence → activation pattern is characterized by the strongest clusters in a region (i.e., the most concentrated clusters in space and time). It is possible in this technique to quantify the strength of a cluster and eliminate the need for apriori regionalization in space and time.

The characteristics of the quiescence → activation patterns, at least for the last three large earthquakes in Southern California, are superficially very similar with periods of quiescence starting approximately 10 years before the large events and periods of activation starting 2 to 3 years prior to the large event. The period of quiescence are characterized by the lowest activity in the regions. Last available data indicates the recent occurrence of a similar pattern in the San Jacinto region. It is suggested that this area deserves attention as a region of presently high seismic risk.

Proposed possible physical explanations for observed clustering (as opposed to computational algorithms) are not

numerous. Examples include inhomogeneities in material strengths along a fault zone ("asperities") and non-linear dynamic friction on fault zones. These were discussed in section 1.11. The asperity model gives a qualitative explanation of the quiescence → activation pattern observed in Chapter 3. Precursory patterns such as quiescence and swarms may reflect the physical properties of fault zones rather than the physics of the failure process.

Chapter 4 applies three proposed precursory seismic patterns termed "bursts of seismicity" to retrospective prediction of large earthquakes which have occurred in the vicinity of the Cocos - North American - Caribbean triple junction. The three patterns, "bursts of aftershocks", swarms, and sigma, are based on abnormal clustering of earthquakes in time, space and energy. The patterns give respectable results although the available data set is not of the highest quality. All occurrences of the patterns are close to the proposed location of the unstable Cocos - North American - Caribbean triple junction. Success of the patterns may be due to the complex nature of the tectonics in the triple junction area. There is no adequate physical explanation for the "bursts of seismicity" patterns.

5.2 Further extensions of this work

There are many gaps in this study. The most obvious gap is a lack of any physical theory which explains the obverved clustering. Any extension of this work should be towards a viable physical model model which offers an explanation of the observations presented here: the self-similarity in the occurrence of aftershocks; the observed clustering of mainshocks; the quiescence → activation pattern observed before many large earthquakes; and the "bursts of seismicity" patterns. A physical model for foreshocks, aftershocks, and mainshocks should be sought. Occurrences of precursory patterns have been independently studied by many investigators, and are well documented. Any model of sequences of earthquakes which fails to explain the well documented patterns such as swarms and quiescence suffers a severe deficiency. A viable model should also have the characteristics of self-similarity.

Cluster analysis appears to be a useful tool for the study of seismicity patterns. The location and study of seismic gaps could be approached using the technique. Another possible use of cluster analysis would be in the development of a statistical model of clustering in space and time. A more physical means of determining the normalization of time with respect to distance is needed as well as a more physical metric for the determination of dissimilarity which incorporates the earthquake magnitudes. However, these improvements require some input from a

theoretical model.

In conclusion it may be worthwhile to speculate on useful directions for study of the physics of the earthquake generating mechanism. The earthquake generating mechanism must in the final analysis be a thermodynamic one. Several constraints must apply to any model of this process.

1. Most of the time an area prone to earthquakes is either in thermodynamic equilibrium or undergoing processes which are thermodynamically (but not mechanically) reversible.
2. The thermodynamic system is non-linear and as a consequence the potentials that describe it (whatever they are) can have more than one stationary point.
3. The system can be characterized by internal and external parameters.
4. Slow changes in the external parameters could change stable equilibrium points to unstable ones and transitions from unstable states to stable states can be earthquakes.
5. In such a non-linear system cluster activity could occur.
6. If this notion is correct models that isolate elements of the process are unlikely to succeed. Perhaps an earthquake is a collective phenomenon.

Perhaps a future research project could address the construction of a macroscopic thermodynamic model, with probabilistic properties, of the earthquake generating

process.

References

Abramowitz, M., and I. A. Segun (1968). *Handbook of Mathmatical Functions*, Dover Publications, New York.

Acharya, H. K. (1979). A method to determine the duration of quiessence in a seismic gap; *Geophys. Res. Lett.* **6**, 681-684.

Aki, K. (1979). Characterization of barriers on an earthquake fault; *J. Geophys. Res.* **84**, 6141-6148.

Allen, C. R. (1975). Geological criteria for evaluating seismicity; *Bull. Geol. Soc. Amer.* **86**, 1041-1057.

Allen, C. R., P. St. Amand, C. F. Richter, and J. M. Nordquist (1965). Relationship between seismicity and geological structure in the southern California region; *Bull. Seismol. Soc. Amer.* **55**, 753-797.

Anderson, J. G. (1979). Estimating the seismicity from geological structure for seismic-risk studies; *Bull. Seismol. Soc. Amer.* **69**, 135-158.

Ando, M. (1975). Source mechanism and tectonic significance of historical earthquakes along the Nankai trough, Japan; *Tectonophysics* **27**, 119-140.

Barenblatt, G. I., V. I. Keilis-Borok, and M. M. Vishik (1981). A model of the clustering of earthquakes; *Geophys. J. Roy. Astron. Soc.* (in press).

Bloom, E. D., and R. C. Erdmann (1979). Frequency - magnitude - time relations in the NGSDC earthquake data file; *Bull. Seismol. Soc. Amer.* **69**, 2085-2099.

Borovik, N., L. Misharina, and A. Treskov (1971). On the possibility of strong earthquakes in Pribakalia in the future; *Izu. Aca. Sci. USSR Phys. Solid Earth* 1, 13-16.

Brady, B. T. (1974). Theory of earthquakes - I. A scale independent theory of rock failure; *Pure and Appl. Geophys.* **112**, 701-725.

Brady, B. T. (1975). Theory of earthquakes - II. Inclusion theory of crustal earthquakes; *Pure and Appl. Geophys.* **113**, 149-168.

Brady, B. T. (1976a). Theory of earthquakes - III. Inclusion theory of deep earthquakes; *Pure and Appl. Geophys.* **114**, 1-21.

Brady, B. T. (1976b). Theory of earthquakes - IV. General implications for earthquake prediction; *Pure and Appl. Geophys.* **114**, 1031-1073.

Brady, B. T. (1976c). Dynamics of fault growth - A physical basis for aftershock sequences; *Pure and Appl. Geophys.* **114**, 727-739.

Brune, J. N. (1968). Seismic moment, seismicity and rate of slip along major fault zones; *J. Geophys. Res.* **73**, 777-784.

Bufe, C. G., P. W. Harsh, and R. O. Burford (1977). Steady-state seismic slip: a precise recurrence model; *Geophys. Res. Lett.* **4**, 91-94.

Bufe, C. G., J. H. Pfluke, and R. L. Wesson (1974). Premonitory vertical migration of micro-earthquakes in Central California - evidence of dilatancy biasing; *Geophys. Res. Lett.* **1**, 221-224.

Byerlee, J. (1978). A review of rock mechanics studies in the United States pertinent to earthquake prediction; *Pure and Appl. Geophys.* **116**, 586-602.

Caputo, M., P. Gasperini, V. keilis-Borok, L. Marcelli, and I. Rotvain (1977). Earthquake swarms as foreunners of strong earthquakes in Italy; *Annali di Geofisica,* *XXX* **N3-4**, 263-283.

Chinnery, M. A. (1969). Earthquake magnitude and source parameters; *Bull. Seismol. Soc. Amer.* **59**, 1969-1982.

Cramer, H. (1946). *Mathematical Methods of Statistics*, Princeton University Press, Princeton.

Das, S., and K. Aki (1977a). A numerical study of two-dimensional spontaneous rupture propagtion; *Geophys. J. Roy. Astron. Soc.* **50**, 643-668.

Das, S., and K. Aki (1977b). Fault plane with barriers: a versatile earthquake model; *J. Geophys. Res.* **82**, 5658-5670.

Davies, G. F., and J. N. Brune (1971). Regional and global fault slip rates from seismicity; *Nature* **229**, 101-107.

Delsemme, J., and A. T. Smith (1979). Spectral analysis of earthquake migration in South America; *Pure and Appl. Geophys.* **117**, 1258-1285.

Dobrovolsky, I. P., S. I. Zubkov, and V. I. Miachkin (1979). Estimation of the size of earthquake preparation zones; *Pure and Appl. Geophys.* **117**, 1025-1044.

Duda, S. (1965). Secular seismic release in the circum-Pacific belt; *Tectonophysics* **2**, 409-452.

Duran, B. S., and P. L. Odell (1974). *Cluster Analysis A Survey*, Springer-Verlag, New York.

Dziewonsky, A., and A. G. Prozorov (1981). Space-time clustering of earthquakes; (in press).

Ellsworth, W. L., A. G. Lindh, W. H. Prescott, and D. G. Herd (1981). The 1906 San Fransisco earthquake and the seismic cycle; (in press).

Engdahl, E. R., and C. Kesslinger (1977). Seismological precursors to a magnitude 5 earthquake in the central Aleutian Islands; *Joun. Phys. Earth* **25**, 5243-5250.

Evernden, J. F. (1970). Study of regional seismicity and associated problems; *Bull. Seismol. Soc. Amer.* **60**,

393-446.

Evernden, J. F. (1971). Location capability of various seismic networks; *Bull. Seismol. Soc. Amer.* **61**, 241-256.

Evison, F. F. (1977a). Fluctuations of seismicity before major earthquakes; *Nature* **266**, 710-712.

Evison, F. F. (1977b). The precursory earthquake swarm; *Phys. Earth Planet. Interiors* **15**, p19-p23.

Evison, F. F. (1977c). Precursory seismic sequences in New Zealand; *N. Z. Joun. of Geol. and Geophys.* **20**, 129-141.

Evison, F. F. (1977c). Long-term seismic precursor to the 1968 Inangahua earthquake, New Zealand; *N. Z. Joun. of Geol. and Geophys.* **21**, 531-534.

Fedotov, S. A. (1965). Regularities of the distribution of strong earthquakes in Kamchatka, the Kurile Islands and northeastern Japan; *Tran. Acad. of Sci. USSR, Inst. Phys. Earth* **36**, 66-93.

Fedotov, S. A., N. A. Dolbilkina, V. N. Morozov, V. I. Myachkin, V. B. Prebrazensky, and G. A. Sobolev (1970). Investigation on earthquake prediction in Kamchatka; *Tectonophysics* **9**, 249-258.

Friedman, M. E., J. H. Whitcomb, C. R. Allen, and J. A. Hileman (1976). *Seismicity of the southern California region, 1 January 1972 to 31 December 1974*, Contribution No. 2734, Division of Geological and Planetary Sciences, California Institute of Technology.

Fuis, G. S., and A. G. Lindh (1979). A change in fault-plane orientation between foreshocks and aftershocks of the Galway Lake earthquake, M=5.2, 1975, Mojave Desert, California (abstract); *Tectonophysics* **52**, 601-602.

Gardner, J. K., and L. Knopoff (1974). Is the sequence of

earthquakes in southern California, with aftershocks removed Poissonian?; *Bull. Seismol. Soc. Amer.* **64**, 1363-1367.

Gasperini, P., M. Caputo, V. I. Keilis-Borok, G. Marcelli, and I. M. Rotvain (1978). A swarms of weak tremors as precursors of strong earthquakes in Italy, in *Vichislitelnaya seismologiya, (Computational seismology), vol. 11*, Nauka Press, Moscow, 3-13.

Gelfand, I. M., Sh. I. Guberman, V. I. Keilis-Borok, and E. Ja. Ranzman (1972). Criteria of high seismicity, determined by pattern recognition; *Tectonophysics* **13**, 415-422.

Gelfand, I. M., Sh. I. Guberman, M. S. Kalezkaja, V. I. Keilis-Borok, E. Ja. Ranzman, I. M. Rotvain, and M. P. Zhidkov (1974). Recognition of places where strong earthquakes may occur, II: Four regions of Asia Minor and S-E Europa, in *Vichislitelnaya seismologiya, (Computational seismology), vol. 7*, Nauka Press, Moscow, 54-71.

Gelfand, I. M., Sh. A. Guberman, V. I. Keilis-Borok, L. Knopoff, F. Press, E. Ya. Ranzman, I. M. Rotwain, and A. M. Sadovsky (1976). Pattern Recognition applied to earthquake epicenters in California; *Phys. Earth Planet. Interiors* 11, 227-283.

Geller, R. J. (1976). Scaling relation for earthquake source parameters and magnitudes; *Bull. Seismol. Soc. Amer.* **66**, 1501-1523.

Gorshkov, A. I., M. Caputo, V. I. Keilis-Borok, E. N. Ofitserova, E Ya. Rantsman, and I. M. Rotvain (1979). Pattern recognition applied to sites of possible occurrence of strong earthquakes. IX Italy, $M \geq 6.0$, in *Vichislitelnaya seismologiya, (Computational seismology), vol. 12*, Nauka Press, Moscow, 3-17.

Gutenberg, B. and C. F. Richter (1954). *Seismicity of the Earth and Associated Phenomena*, Princeton University Press, New Jersey.

Hasegawa, A., T. Hasegawa, and S. Hori (1975). Premonitory

variation in seismic velocity related to the Southeastern Akita earthquake of 1970; *Joun. Phys. Earth* **23**, 189-203.

Hileman, J. A., C. R. Allen, and J. M. Nordquist (1974). *Seismicity of the southern California region, 1 January 1932 to 31 December 1972*, Contribution No. 2385, Division of Geological and Planetary Sciences, California Institute of Technology.

Holmes, R. A., and R. B. MacDonald (1969). The physical basis of systems design for remote sensing in agriculture; *Proceedings IEEE* **57**, 629-639.

Imamura, A. (1928). On the seismic activity of central Japan; *Jap. J. Astron. Geophys.* **6**, 119-137.

Inouye, W. (1965). On the seismicity in the epicentral region and its neighborhood before the Niigata earthquake; *Kenshin Jiho* **29**, 31-36.

Ishida, M., and H., Kanamori (1977). The spatio-temporal variation of seismicity before the 1971 San Fernando earthquake, California; *Geophys. Res. Lett.* **4**, 345-346.

Ishida, M., and H. Kanamori (1978). The foreshock activity of the 1971 San Fernando earthquake, California; *Bull. Seismol. Soc. Amer.* **68**, 1265-1279.

Ishida, M., and H. Kanamori (1980). Temporal variation of seismicity and spectrum of small earthquakes preceding the 1952 Kern County, California earthquake; *Bull. Seismol. Soc. Amer.* **70**, 509-527.

Ito, K. (1980). Periodicity and chaos in great earthquake occurrence; *J. Geophys. Res.* **85**, 1399-1408.

Jardine, N., and R. Sibson (1971). *Mathematical Taxonomy*, John Wiley & Sons Ltd., London.

Jones, L. M., and P. Molnar (1976). Frequency of foreshocks; *Nature* **262**, 677-679.

Jones, L. M., and P. Molnar (1979). Some characteristics of foreshocks and their possible relationship to earthquake prediction and premonitory slip on faults; *J. Geophys. Res.* **84**, 3596-3608.

Kagan, Y., and L. Knopoff (1976). Statistical search for non-random features of the seismicity of strong earthquakes; *Phys. Earth Planet. Interiors* **12**, 291-318.

Kagan, Y., and L. Knopoff (1977). Earthquake risk prediction as a stochastic process; *Phys. Earth Planet. Interiors* **14**, 97-108.

Kagan, Y., and L. Knopoff (1978). Statistical study of the occurrence of shallow earthquakes; *Geophys. J. Roy. Astron. Soc.* **55**, 67-86.

Kagan, Y., and L. Knopoff (1980a). Spatial distribution of earthquakes: the two-point correlation function; *Geophys. J. Roy. Astron. Soc.* **62**, 303-320.

Kagan, Y., and L. Knopoff (1980b). Dependence of seismicity on depth; *Bull. Seismol. Soc. Amer.* **70**, 1811-1822.

Kanamori, H. (1977a). Seismic and aseismic slip along subduction zones and their tectonic implications, in *Island Arcs, Deep Sea Trenches and Back-Arc Basins*, AGU, Washington, 163-174.

Kanamori, H. (1977b). The energy release in great earthquakes; *J. Geophys. Res.* **82**, 2981-2987.

Kanamori, H. (1981). The nature of seismicity patterns before large earthquakes; *Maurice Ewing Series, 4, AGU* (in press).

Kanamori, H., and K. Abe (1979). Reevaluation of the turn-of-the-century seismicity peak; *J. Geophys. Res.* **84**, 6131-6139.

Kanamori, H., and D. L. Anderson (1975). Theoretical basis of some empirical relations in seismology; *Bull. Seismol. Soc. Amer.* **65**, 1073-1095.

Kanamori, H., and G. S. Stewart (1978). Seismological aspects of the Guatemala earthquake of February 21, 1976; *J. Geophys. Res.* 83, 3427-3434.

Kantorovich, L. V., G. M. Molchan, V. I. Keilis-Borok, and E. V. Vil'kovich (1970). A statistical model of seismicity and an an estimate of the basic seismic effects; *Izu. Aca. Sci. USSR Phys. Solid Earth* 5, 85-102.

Kasahara, K. (1973). Earthquake fault studies in Japan; *Phil. Trans. R. Soc. Lond.* **274**, 287-296.

Keilis-Borok, V. I., V. M. Podgayetskaya, and A. G. Prosorov (1971). On the local statistics of catalogs of earthquakes, in *Vichislitelnaya seismologiya, (Computational seismology), vol. 5*, Nauka Press, Moscow, 55-79.

Keilis-Borok, V. I., L. Knopoff, and I. M. Rotvain (1980a). Bursts of aftershocks, long-term precursors of strong earthquakes; *Nature* **283**, 259-263.

Keilis-Borok, V. I., L. Knopoff, I. M. Rotvain, and T. M. Sidorenko (1980b). Bursts of seismicity as long-term precursors of strong earthquakes; *J. Geophys. Res.* **85**, 803-811.

Keilis-Borok, V. I., L. Knopoff, and C. R. Allen (1980c). Long-term premonitory seismicity patterns in Tibet and the Himalayas; *J. Geophys. Res.* **85**, 813-820.

Keilis-Borok, V. I., and L. N. Malinovskaya (1964). One regularity in the occurrence of strong earthquakes; *J. Geophys. Res.* **69**, 3019-3024.

Keilis-Borok, V. I., V. M. Podgayetskoya, and A. G. Prozoroff (1971). On the local statistics of catalogs of earthquakes; *Computational Seismology* 5, 55-79.

Keilis-Borok, V. I., and I. M. Rotvain (1978). On soon-to-break seismic gaps, in *Proc. Conf. VI: Methodology for Identifying Seismic Gaps and Soon-to-break Gaps*, U. S. Geol. Surv., Open-File

Report 78-943, 387-394.

Keilis-Borok, V. I., and I. M. Rotvain (1979). Two long-range precursors of strong earthquakes, in *Vichislitelnaya seismologiya, (Computational seismology), vol. 12,* Nauka Press, Moscow, 18-27.

Kelleher, J. A. (1970). Space-time seismicity of the Alaska-Aleutian seismic zone; *J. Geophys. Res.* 75, 5745-5756.

Kelleher, J. A. (1972). Rupture zones of large South American earthquakes and some predictions; *J. Geophys. Res.* 77, 2087-2103.

Kelleher, J. A., and W. McCann (1976). Bouyant zones, great earthquakes, and unstable boundaries of subduction; *J. Geophys. Res.* 81, 4885-4896.

Kelleher, J. A., and W. McCann (1977). Bathymetric highs and the development of convergent plate boundaries, in *Island Arcs, Deep Sea Trenches, and Back-Arc Basins,* AGU, Washington, 99-114.

Kelleher, J., L. Sykes, and J. Oliver (1973). Possible criteria for predicting earthquake locations and their application to major plate boundaries of the Pacific and Caribbean; *J. Geophys. Res.* 78, 2547-2585.

Kelleher, J. A., and J. Savino (1975). Distribution of seismicity before large strike slip and thrust-type earthquakes; *J. Geophys. Res.* 80, 260-271.

Kelleher, J. A., J. Savino, H. Rowlett, and W. McCann (1974). Why and where great thrust earthquakes occur along island arcs; *J. Geophys. Res.* 79, 4889-4899.

Kelleher, J., L. K. Sykes, and J. Oliver (1973). Possible criteria for predicting earthquake locations and their application to major plate boundaries of the Pacific and Caribbean; *J. Geophys. Res.* 78, 2547-2585.

Khattri, K., and M. Wyss (1978). Precursory variation of seismicity rate in the Assam area, India; *Geology* 6, 685-688.

Knopoff, L. (1971). A stochastic model for the occurrence of Main-Sequence earthquakes; *Rev. Geophys. Space Phys.* 9, 175-188.

Knopoff, L., and Y. Kagan (1977). Analysis of the theory of extremes as applied to earthquake problems; *J. Geophys. Res.* 82, 5647-5657.

Kosobokov, V., and I. Rotvain (1977). Pattern recognition of earthquake prone areas, VI. magnitude $M \geq 7$, in *Vichislitelnaya seismologiya, (Computational seismology), vol. 10*, Nauka Press, Moscow, 88-105.

Kristy, M. J., and D. W. Simpson (1980). Seismicity changes preceding two recent central Asian earthquakes; *J. Geophys. Res.* 85, 4829-4837.

Landgrebe, D. A., and LARS Staff (1969). *LARSYAA, a processing system for airborne earth resource data*, LARS Information Note 091968, Purdue University, Layfayette, Indiana.

Landgrebe, D. A., and T. L. Phillips (1967). A multichannel image data handling system for agriculture remote sensing, in *Proc. Seminar on Computerized Image Handling Techniques*, Washington, D. C., XIT 1-10.

Lay, T., and H. Kanamori (1980). Earthquake doublets in the Solomon Islands; *Phys. Earth Planet. Interiors* 21, 283-304.

Lukk, A., and V. Ponomarev (1972). Trends in the variation of the seismic background in the course of time; *Izu. Aca. Sci. USSR Phys. Solid Earth* 5, 3, 271-275.

McEvilly, T., W. Bakun, and K. Casady (1967). The Parkfield, California earthquake of 1966 Cocos plate; *Bull. Seismol. Soc. Amer.* 57, 1221.

McCann, W. R., S. P. Nishenko, L. R. Sykes, and J. Kraus

(1979). Seismic gaps and plate tectonic: Seismic potential for major boundaries; *Pure and Appl. Geophys.* **117**, 1082-1147.

McNally, K. C. (1977). Patterns of earthquake clustering preceding moderate earthquakes, central and southern California, abstract; *EOS* **58**, 1195.

McNally, K. C. (1981). Plate subduction and prediction of earthquakes along the Middle America trench; (in press).

McNally, K. C., and J. B. Minster (1981). Non-uniform seismic slip rates along the Middle America Trench; *J. Geophys. Res.* **86**, 4949-4959.

Minster, J. B., and T. H. Jordan (1978). Present-Day plate motions; *J. Geophys. Res.* **83**, 5331-5354.

Mizone, M., M. Nakamura, Y. Ishiketa, and N. Seta (1978). Earthquake prediction from micro-earthquake observation in the vicinity of Wakayama city, Northwestern part of the Kii peninsula, central Japan; *Joun. Phys. Earth* **26**, 397-416.

Mogi, K. (1968a). Some features of recent seismic activity in and near Japan, 1; *Bull. Earthquake Res. Inst. Tokyo* **46**, 1225-1236.

Mogi, K. (1968b). Sequential occurrences of recent great earthquakes; *Joun. Phys. Earth* **16**, 30-36.

Mogi, K. (1968c). Migration of seismic activity; *Bull. Earthquake Res. Inst. Tokyo* **46**, 53-74.

Mogi, K. (1969a). Some features of recent seismic activity in and near Japan, 2, Activity before and after great earthquakes; *Bull. Earthquake Res. Inst. Tokyo* **47**, 395-417.

Mogi, K. (1969b). Relationship between the occurrence of great earthquakes and tectonic structure; *Bull. Earthquake Res. Inst. Tokyo* **47**, 429-451.

Mogi, K. (1979). Two kinds of Gaps; *Pure and Appl. Geophys.* **117**, 1172-1186.

Molnar, P. (1979). Earthquake recurrence intervals and plate tectonics; *Bull. Seismol. Soc. Amer.* **69**, 115-133.

Nur, A. (1978). Nonuniform friction as a physical basis for earthquake mechanics; *Pure and Appl. Geophys.* **116**, 964-991.

Ohtake, M. T. (1976). Search for precursors of the 1974 Izu-Hanto-Oki earthquake, Japan; *Pure and Appl. Geophys.* **114**, 1083-1093.

Ohtake, M., T. Matumoto and G. V. Lathan (1977). Seismicity gap near Oaxaca, Southern Mexico as a probable precursor to a large earthquake; *Pure and Appl. Geophys.* **115**, 375-385.

Papazachos, B. (1974). On certain aftershock and foreshock parameters in the area of Greece; *Ann. Geofis.* **27**, 497-515.

Papazachos, B. (1975). Foreshocks and earthquake prediction; *Tectonophysics* **28**, 213-226.

Papazachos, B., M. Delibasis, N. Liapis, G. Moumoulidis, and G. Purcaru (1967). Aftershock sequences of some large earthquakes in the region of Greece; *Ann. Geofis.* **20**, 1-93.

Powell, J. A., and S. J. Duda (1975a). A statistical study of earthquake occurrence; *Pure and Appl. Geophys.* 113, 447-460.

Powell, J. A., and S. J. Duda (1975b). An earthquake migration pattern of high statistical significance; *Pure and Appl. Geophys.* **113**, 461-466.

Prozorov, A. G., and A. Dziewonski (1981). A method of studying variations in the clustering properties of earthquakes: application to the analysis of global seismicity; (in press).

Quittmeyer, R. C. (1979). Seismicity variations in the Makran region of Pakistan and Iran: relation to great earthquakes; *Pure and Appl. Geophys.* **117**, 1212-1228.

Rikatake, T. (1974). Probability of earthquake occurrence as estimated from crustal strain; *Tectonophysics* **23**, 299-312.

Rikitake, T. (1975a). Statistics of ultimate strain of the earths crust and probability of earthquake occurrence; *Tectonophysics* **26**, 1-21.

Rikitake, T. (1975b). Earthquake precursors; *Bull. Seismol. Soc. Amer.* **65**, 1133-1162.

Rikitake, T. (1976a). Recurrence of great earthquakes at subduction zones; *Tectonophysics* **35**, 335-362.

Rikitake, T. (1976b). *Earthquake Prediction*, Elsevier, Amsterdam.

Rikitake, T. (1979). Classification of earthquake precursors; *Tectonophysics* **54**, 293-309.

Sadovsky, M., I. Nersesov, S. Nigmatullaev, L. Latynina, A. Lukk, A. Semenov, I. Sembireva, and V. Ulomov (1972). The process preceding strong earthquakes in some regions of middle Asia; *Tectonophysics* **14**, 295-307.

Sampson, R. J. (1975). *Surface II Graphics System*, Kansas Geological Survey, Lawrence, Kansas.

Sauber, J., and P. Talwani (1980). Application of Keilis-Borok and McNally algorithms to earthquakes in the Lake Jocassee area, South Carolina; *Phys. Earth Planet. Interiors* **21**, 267-281.

Savage, J. C., and W. H. Prescott (1978). Asthenosphere readjustment and the earthquake cycle; *J. Geophys. Res.* **83**, 3369-3381.

Scholz, C. H. (1977). A physical interpretation of the Haicheng earthquake prediction; *Nature* **267**, 121-124.

Scholz, C. H., L. R. Sykes, and Y. P. Aggarwal (1973). Earthquake prediction: a physical basis; *Science* **181**, 803-810.

Sekiya, H. (1977). Anomalous seismic activity and earthquake prediction; *Joun. Phys. Earth* **25**, 585-593.

Shimazaki, K., and T. Nakata (1980). Time-predictable recurrence model for large earthquakes; *Geophys. Res. Lett.* 7, 279-282.

Sieh, K. (1978). Prehistoric large earthquakes produced by slip on the San Andreas fault at Pallett Creek, California; *J. Geophys. Res.* **83**, 3907-3948.

Singh, S. K., L. Astiz, and J. Havskov (1981). Seismic gaps and recurrence periods of large earthquakes along the Mexican subduction zone: a reexamination; *Bull. Seismol. Soc. Amer.* 71, 827-843.

Späth, H. (1980). *Cluster Analysis Algorithms for Data Reduction and Classification of Objects*, Ellis Horwood Publishers, Chichester, England.

Suyehiro, S. (1966). Difference between aftershocks and foreshocks in the relationship of magnitude to frequency of occurrence for the great Chilean earthquake of 1960; *Bull. Seismol. Soc. Amer.* **56**, 185-200.

Suyehiro, S., and H. Sekiya (1972). Foreshocks and earthquake prediction; *Tectonophysics* **14**, 219-225.

Sykes, L. R. (1971). Afershock zones of great earthquakes, seismicity gaps, and earthquake prediction for Alaska and the Aleutians; *J. Geophys. Res.* **76**, 8021-8041.

Sykes, L. R., and R. C. Quittmeyer (1981). Repeat times of great earthquakes along simple plate boundaries; (in press).

Talwani, p. (1979). An empirical earthquake prediction model; *Phys. Earth Planet. Interiors* 18, 288-302.

Thatcher, W., J. A. Hileman, and T. C. Hanks (1977). Seismic slip distribution along the San Jacinto fault zone, Southern California, and its implications; *Bull. Geol. Soc. Amer.* 86, 1140-1146.

Toksöz, M. N., A. F. Shakal, and A. J. Michael (1979). Space-time migration of earthquakes along the north Anatolian fault zone and seismic gaps; *Pure and Appl. Geophys.* 117, 1258-1270.

Utsu, T. (1972). Aftershocks and earthquake statistics; *J. Fac. Sci. Hokkaido Univ. Ser. VII, Geophysics* N3, 379-441.

Utsu, T. (1972). Aftershocks and earthquake statistics; *J. Fac. Sci. Hokkaido Univ. Ser. VII, Geophysics* N4, 1-42.

Utsu, T. (1974). Space-time pattern of large earthquakes occurring off the Pacific coast of the Japanese Islands; *Joun. Phys. Earth* 22, 325-342.

Vere-Jones, D. (1975). Stocastic models for earthquake sequences; *Geophys. J. Roy. Astron. Soc.* 42, 811-826.

Vere-Jones, D. (1976). A branching model for crack propagation; *Pure and Appl. Geophys.* 114, 711-725.

Vere-Jones, D. (1978). Earthquake prediction: A statistician's view; *Joun. Phys. Earth* 26, 129-146.

Wei, G., Z. Lin, X. Zhu, Y. Zhao, X. Zhao, and H. Hou (1978). On seismic gaps previous to certain great earthquakes occurring in North China; *Acta Geophysica Sinica* 21, 213-217.

Wesson, R. L., and W. L. Ellsworth (1973). Seismicity preceding moderate earthquakes in California; *J. Geophys. Res.* 78, 8527-8543.

Whitcomb, J. H., J. D. Garmany, and D. L. Anderson (1973). Earthquake prediction: variation of seismic velocities before the San Fernando earthquake; *Nature* **181**, 632-635.

Wishart, D. (1969). Numerical classification method for deriving natural classes; *Nature* **221**, 97-98.

Wishart, D. (1978). *CLUSTAN User Manual (third edition)*, Program Library Unit, Edinburgh University, Edinburgh, Scotland.

Wood, M. D., and S. S. Allen (1973). Recurrence of seismic migration along the central California segment of the San Andreas fault system; *Nature* **244**, 213-215.

Wu, K., M. Yue, H. Wu, X. Cao, H. Chen, W. Haung, K. Tain, and S. Lu (1976). Certain characteristics of the Haicheng earthquake (M = 7.3) sequence; *Acta Geophys. Sinica* **19**, 109-117.

Wyss, M. (1973). Towards a physical understanding of the earthquake frequency distribution; *Geophys. J. Roy. Astron. Soc.* **31**, 341-360.

Wyss, M., and R. E. Habermann (1979). Seismic quiescence precursory to a past and a future Kurile Island earthquake; *Pure and Appl. Geophys.* **117**, 1195-1211.

Wyss, M., and W. H. K. Lee (1973). Time variations of the average earthquake magnitude in Central California; *Proc. Conf. Tecton. Prob. San Andreas Fault System, Stanford Univ. Publ.* **13**, 24-42.

Yamashina, K., and Y. Inoue (1979). A doughnut-shaped pattern of seismic activity preceding the Shimane earthquake of 1978; *Nature* **278**, 48-50.

Yamashina, K. (1981). Some empirical rules on foreshocks and earthquake prediction; (in press).

Yegulalp, T. M., and J. T. Kuo (1974). Statistical prediction of the occurrence of maximum magnitude earthquakes; *Bull. Seismol. Soc. Amer.* **64**, 393-414.